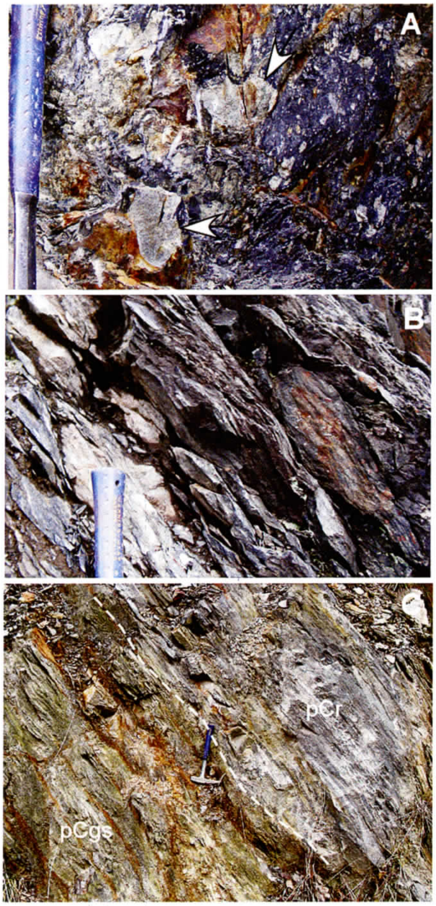
# Southeastern Geology: Volume 45, No. 4
# August 2008

Editor in Chief: S. Duncan Heron, Jr.

**Abstract**

Academic journal published quarterly by the Department of Geology, Duke University.

# SOUTHEASTERN GEOLOGY




Vol. 45, No. 4

August 2008

# SOUTHEASTERN GEOLOGY

PUBLISHED

at

DUKE UNIVERSITY

Duncan Heron

Editor in Chief

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Editor


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**SOUTHEASTERN GEOLOGY** is a peer review journal.

ISSN 0038-3678

# SOUTHEASTERN GEOLOGY

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# RE-INTERPRETATION OF THE DEFORMATION HISTORY OF THE GREENBRIER FAULT, GREAT SMOKY MOUNTAINS: CRITICAL ASSESSMENT OF PREVIOUS WORK

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## ABSTRACT

**The Greenbrier fault as originally defined fails to explain many aspects of Great Smoky and Snowbird group geology and introduces unnecessary complexities for the tectonic history of the Great Smoky Mountains region (e.g. multiple depositional basins). Early workers admitted to limitations in assumptions that were the basis for their interpretation that the fault is premetamorphic and accommodates at least 23 km of shortening. Although original work did not describe a specific exposure of the fault at the type locality, displacement along the Great Smoky-Snowbird contact is inferred from truncation of bedding in rocks adjacent to the contact. This truncation is the only observation consistent with premetamorphic displacement along the contact, but it is also consistent with post-metamorphic slip. Taken as a whole, the evidence permits the alternative interpretation that the Greenbrier is a faulted stratigraphic contact exhibiting mostly postmetamorphic slip between thick units of markedly different competency that does not require tens of kilometers of displacement or deposition in separate basins.**

## INTRODUCTION

The Blue Ridge province of the southern Appalachian Mountains (Figure 1) is a composite terrane dominated by crystalline thrust sheets. The Western Blue Ridge includes Grenville basement, and predominantly thick-bedded metasandstones, interbedded metasiltstones, slates, and phyllites of the Ocoee Supergroup (King et al., 1958; Hadley and Goldsmith, 1963; King, 1964). The contact between the two thickest Ocoee units (Great Smoky and Snowbird Groups) was defined as the Greenbrier fault (King et al., 1958; Hadley and Goldsmith, 1963), an inferred premetamorphic thrust fault with an estimated minimum displacement of 23 km that emplaces presumed younger Great Smoky on older Snowbird rocks. The Greenbrier fault figures prominently in models of southern Appalachian orogenesis, and is widely accepted as a large displacement thrust (e.g., Rast and Kohles, 1986; Connelly and Woodward, 1992). Folding of Snowbird group footwall rocks and thrust faulting were interpreted to have preceded high-grade Taconian regional metamorphism and folding. However, regional metamorphism and locally intense post-metamorphic deformation obscure original relationships and preclude unequivocal interpretation of the nature of the fault. Additional complications relate to uncertainty in stratigraphic assignments and relationships between units along the contact.

Although clearly a lithologic discontinuity of some sort, the interpretation of the Greenbrier

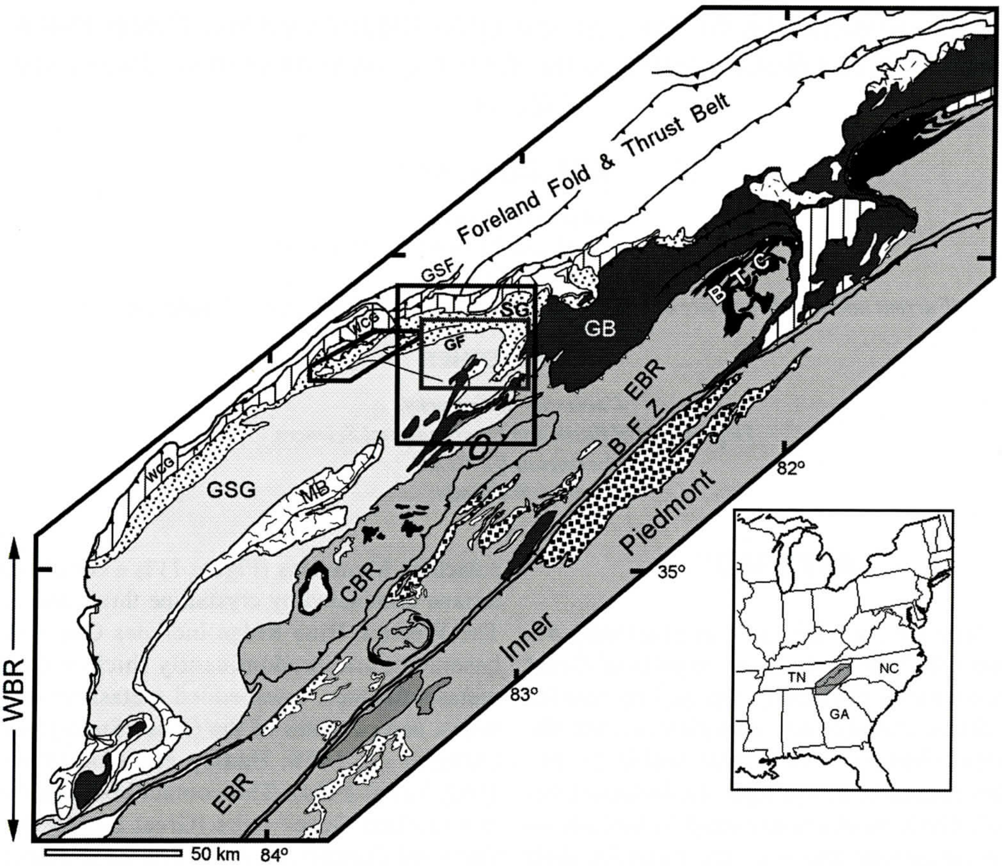



**Figure 1. Regional terrane map of the southern Appalachians showing location (inset) and geologic setting of the study area (after Rankin et al., 1990). Boxes show locations of Figures 2A (gray line), 2B (polygon), and 3 (heavy black line). BFZ: Brevard fault zone; BTC: area of the Blue Ridge thrust complex; CBR: Central Blue Ridge terrane; EBR: Eastern Blue Ridge terrane; GB: Grenville basement; GF: Greenbrier fault; GSF: Great Smoky fault; GSG: Great Smoky group; MB: Murphy belt; SG: Snowbird group; WBR: Western Blue Ridge.**

as a major premetamorphic thrust fault is not entirely consistent with the character of the Snowbird-Great Smoky contact. Nor is it consistent with other aspects of the regional geology. Numerous questions and testable implications arise from the inference that the Great Smoky-Snowbird group contact is a younger-over-older thrust fault with tens of kilometers of offset (e.g., Montes and Hatcher, 1999). The objective of this paper is to summarize salient observations, assumptions, and inferences of the original (Hamilton, 1961; Hadley and Goldsmith, 1963; King, 1964) and subsequent research (Woodward et al., 1991; Southworth et al., 2005a, b) bearing on the interpretation of the contact. We also employ geochemical analysis and examine field and petrologic relationships to test the hypothesis of previous workers that the Metcalf phyllite is the tectonized equivalent of the Pigeon siltstone, which is critical to deciphering the deformation history of the Greenbrier fault. Clemons and Moecher (in subm., Geological Society of America Bulletin [GSAB]) present new structural and petrologic observations from Great Smoky and Snowbird group units along the contact from lower greenschist to upper amphibolite facies, documenting petrologic, petrofab-

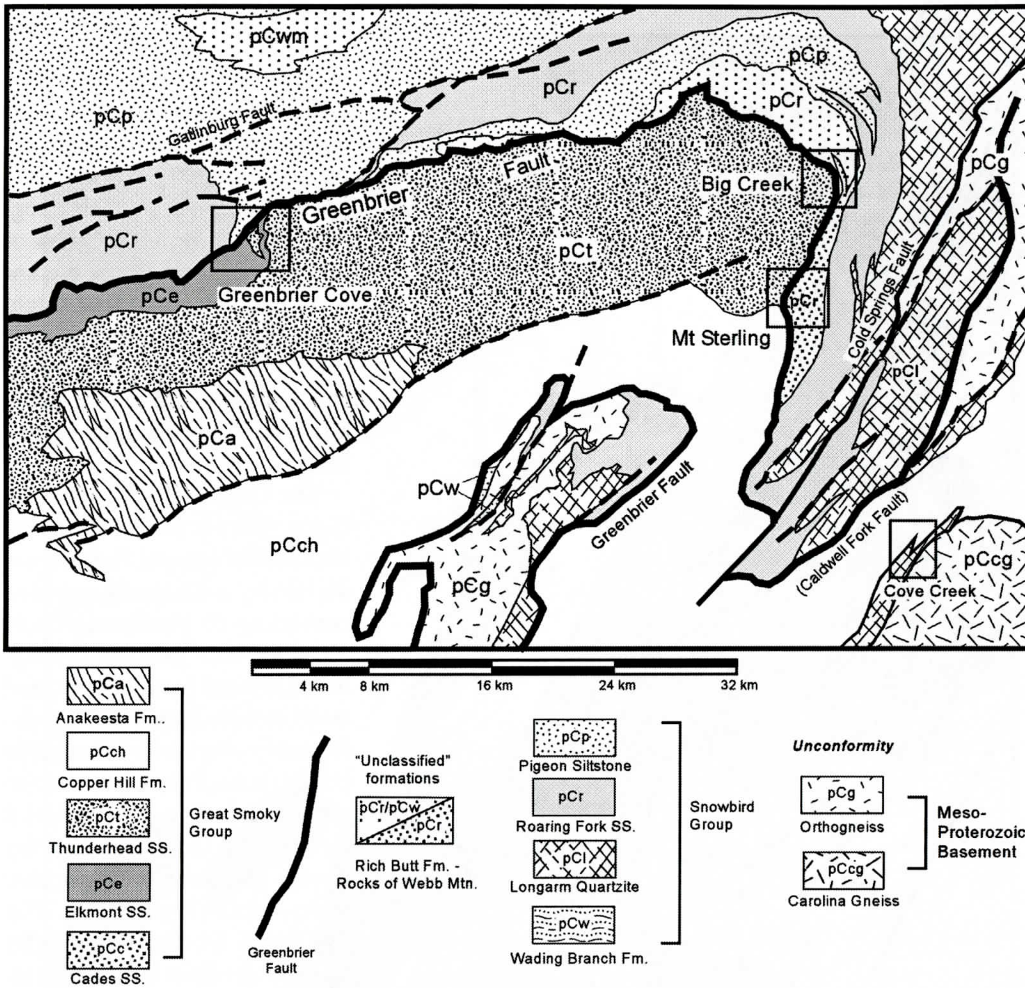
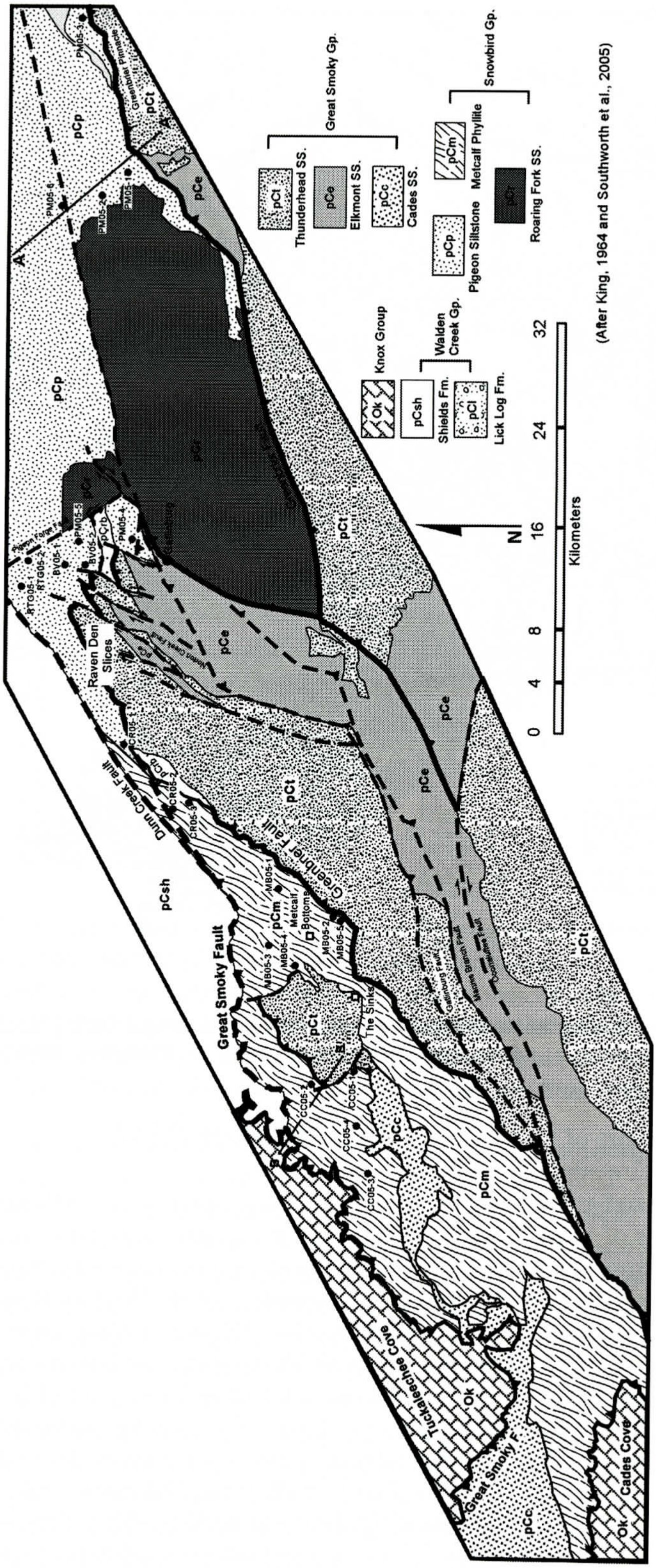



**Figure 2 A. Schematic geologic map of the eastern Great Smoky Mountains area (Hadley and Goldsmith, 1963; Southworth et al., 2005b) with areas of detailed study of Clemons (2006) (boxes).**

ric and field relationships in support of an alternative interpretation that the Greenbrier fault is a post-metamorphic faulted stratigraphic contact that does not require tens of km of displacement. The new data and interpretations bear on relationships of Great Smoky geology to the Blue Ridge thrust complex to the northeast (Goldberg et al., 1992; Trupe et al., 2003; Merschat et al., 2006), the central and eastern Blue Ridge to the south (Hatcher, 2002), and the Blue Ridge of Georgia and Alabama to the southwest (Tull, 2002).

## GEOLOGIC SETTING

The geology and physiography of the Great Smoky Mountains are dominated by the remarkably thick and erosionally resistant Ocoee supergroup, a component of the Western Blue Ridge (WBR) province (Figure 1). Neoproterozoic extension of Mesoproterozoic Laurentian continental crust resulted in formation and infilling of a rift basin or basins by immature quartzofeldspathic clastic sediments derived from rift flanks (Hadley and Goldsmith, 1963; Rankin, 1975; Rast and Kohles, 1986). Subsequent deformation and sub-greenschist to upper




amphibolite facies metamorphism produced a range of deformation styles that make it difficult to unequivocally define the nature of the contact between the Great Smoky and Snowbird groups. Steep terrane, thick vegetation, and the remoteness of this vast region hamper a complete understanding that might arise from detailed quadrangle-scale mapping, and structural, stratigraphic, and petrographic analysis (Southworth et al., 2005b).

The Great Smoky highlands in the vicinity of Great Smoky Mountains National Park are underlain by a basement complex consisting of Mesoproterozoic (Grenvillian) granitic orthogneiss, layered paragneisses, and amphibolite, which are unconformably overlain by fossil-free clastic rocks of the Ocoee Supergroup, including the Snowbird Group and the Great Smoky Group and their metamorphic equivalents (Figure 2A). The Snowbird Group is generally fine- to medium-grained feldspathic metasiltstone, metasandstone, rare metaconglomerate, slate, and phyllite. The Great Smoky group includes feldspathic meta- sandstones and meta-conglomerates (schists and gneisses to the southeast), and shale interbeds (now slate or schist). The contact between the

**Figure 2B. Geologic map of the western and central Great Smoky Mountains (after Hadley and Goldsmith, 1963; King, 1964; and Southworth et al., 2005b), with locations of cross sections shown in Figure 6 and other locations referred to in text.**

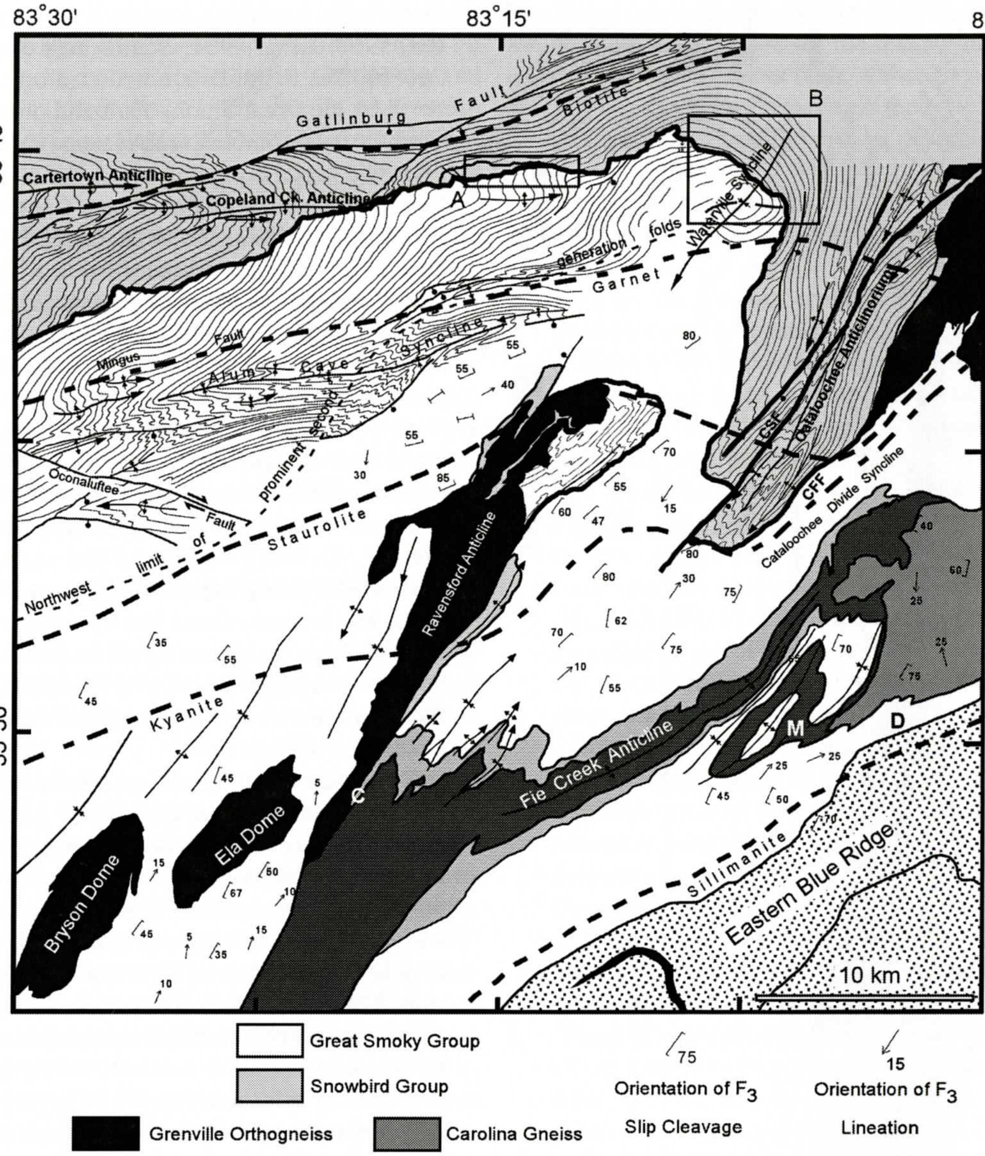
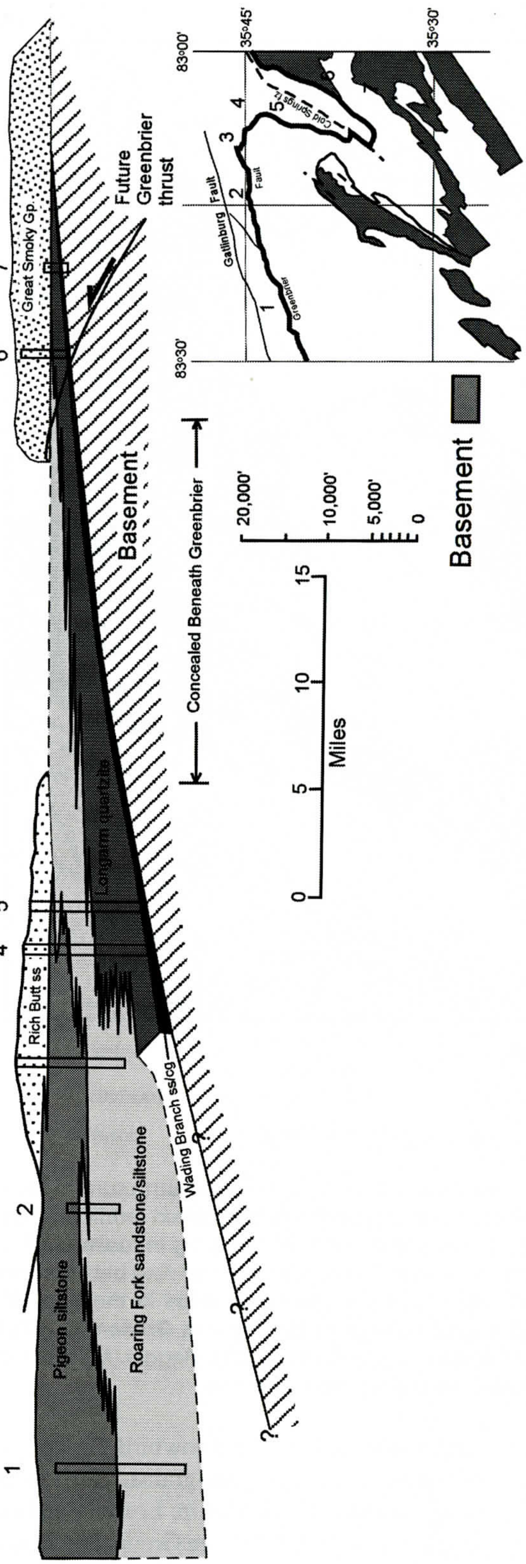



**Figure 3. Plate 3 redrawn directly from Hadley and Goldsmith (1963) showing structural features (major fold trends, faults, and bedding form lines) and locations discussed in text (additional data from Hadley and Nelson, 1971). Area of marked discordance between bedding in footwall (Roaring Fork Fm.) and Greenbrier fault (box A) and area of near discordance of bedding between footwall (Rich Butt Fm.) and fault (box B) are indicated. Northeast-trending folds in the Snowbird group in the area of Cataloochee Anticlinorium (terminology of Hadley and Goldsmith, 1963), folds in the Great Smoky-Snowbird contact in Cherokee (C), Soco Gap (SG), Maggie (M), and Dellwood (D) areas, and major anticlines, are $F_3$ folds; cleavages are $S_3$ fabrics.**

Snowbird and Great Smoky groups is the Greenbrier fault (King et al., 1958; Hadley and Goldsmith, 1963).

Foothills Belt strata of the Great Smoky Mountains include the Snowbird Group, predominantly fine- to coarse-grained clastic rocks of the Walden Creek Group, Lower Cambrian sandstone of the Chilhowee Group, and Middle




Ordovician Jonesboro Limestone (Hamilton, 1961; King, 1964; Southworth et al. 2005b). The Foothills are bounded on the north by the Great Smoky fault and on the south by the Gatlinburg fault system (Figure 2B). The Metcalf phyllite of the Snowbird Group is exposed in a northeast-trending belt from Gatlinburg, TN to the Cades Cove area (Figure 2B). It is bounded below by the Great Smoky or the Dunn Creek fault, and is overlain by Great Smoky group formations along the Greenbrier fault. The Wading Branch, Longarm, Pigeon, and Roaring Fork Formations (also Snowbird Group), occur from the Gatlinburg area to the northeast around the Waterville syncline and then southward toward the Dellwood area (Figures 2A, 3). Interpreted lateral and vertical stratigraphic relationships among Snowbird units, and between Great Smoky (undivided) and Snowbird strata, based on sections described along the length of the Greenbrier Fault, are shown in Figure 4 (Hadley and Goldsmith, 1963).

Regional Barrovian metamorphism ranges from greenschist facies (chlorite grade) north of the Greenbrier fault to upper amphibolite facies (sillimanite-grade migmatization) in the southern Great Smoky Mountain region (Figure 3). Isograds are most likely Middle Ordovician (i.e., Taconian; Moecher et al. 2004, 2005, 2007; Kunk et al., 2006).

Premetamorphic deformation (map-scale folding of Greenbrier footwall units) and peak metamorphism (formation of isograds) were followed by periods of retrograde deformation and mineral growth (Hadley and Goldsmith, 1963; King, 1964; Southworth et al., 2005b; Massey and Moecher, 2005; Moecher et al., 2005, 2007). Post-metamorphic structural features include: (1) map scale folds (Ela and Bryson domes, Ravens-

**Figure 4. Interpreted stratigraphic relationships among Snowbird Group formations, Great Smoky Group (undivided), and basement (after Figure 10, Hadley and Goldsmith, 1963). Great Smoky is conformable with Snowbird Group near Cove Creek (sections 6 and 7).**

ford and Fie Creek anticlines, Walker Bald, Buck Mtn., and Waterville synclines: Figure 3); (2) post-metamorphic faults (Gatlinburg and Oconaluftee faults); (3) ductile high strain zones (e.g., Caldwell Fork and Cold Springs faults and Little River and Eagle Creek Shear zones; Montes, 1997; Montes and Hatcher, 1999; Southworth et al., 2005b; Merschat et al., 2006); (4) regions of distributed post-metamorphic shear (Massey and Moecher, 2005); and meso- to microscopic-scale, tight to isoclinal folding of schistosity and gneissosity with formation of a steep axial planar foliation that overprints slaty cleavage to migmatitic layering ("slip cleavage" of Hadley and Goldsmith, 1963; Massey and Moecher, 2005; Moecher et al., 2005, 2007).

## PREVIOUS RESEARCH

Much of the generally accepted interpretation of the Great Smoky Mountains is based on 1:24,000 to 1:62,500 scale mapping of the Great Smoky Mountains Region by the U.S.G.S. led by P.B. King during and following World War II. This work was a remarkable accomplishment considering the vastness and remoteness of the Great Smoky Mountains region. However, that work was completed over a half century ago. In re-assessing previous research we generally do not question the geologic maps, although Southworth et al. (2005b) and Merschat et al. (2006), present alternative map interpretations. Instead, we submit that re-interpretation of the geologic history of the Great Smoky Mountains region is warranted in light of new field observations and modern developments in petrotectonic and kinematic analysis. Where appropriate, the original observations and inferences of earlier workers are directly quoted to most accurately illuminate their thinking at the time, followed immediately by our critical assessment.

### U.S.G.S. Mapping: Hadley and Goldsmith (1963)

Hadley and Goldsmith (1963) mapped the largest area of the King-U.S.G.S. Great Smoky Mountains survey and provide the most detailed summary of the evidence for the contact they interpret as the Greenbrier fault. Although specific exposures of the contact near the type locality (Greenbrier Pinnacle) are not described, Hadley and Goldsmith (1963) inferred that the contact is a premetamorphic thrust fault (the Greenbrier fault) based primarily on the following observations:

1. Locally, the contact crosscuts strike of bedding along the contact, and is projected at depth to cut bedding, several lithologic contacts, and folds in both the footwall (Snowbird Group) or hanging wall (Great Smoky Group), primarily along the northern mountain front between Greenbrier Pinnacle and Cosby Creek, and along the eastern contact near Mt. Sterling southward into the Cataloochee anticlinorium (Figs. 2A, 2B, 3). Tight to isoclinal folding, attenuation of strata, retrograde ductile shearing, and staurolite to kyanite grade metamorphism south of the Big Creek area (Figure 2A) obliterate evidence for the original nature of the contact.
2. Regional trends of slaty cleavage and metamorphic isograds do not appear to be deflected by the contact (Figure 3). "*Slaty cleavage is not disturbed close to the fault as it certainly would be if movement occurred after metamorphism.*" (Hadley and Goldsmith, 1963, p. B106)
3. There is a significant change in thickness of the Longarm Fm. from approximately 1000 to 4000 feet across the presumed position of the Greenbrier fault on the southeast flank of the Cataloochee anticlinorium. (Figs. 2A, 3). The difference is accounted for by overriding of the missing section by the Greenbrier fault hanging wall (Figure 4). Assuming original lateral continuity and constant tapering angle of the Longarm Fm., the amount of shortening is conservatively estimated to be 15 mi (~23 km). Hadley and Goldsmith (1963) stated:

   "*This estimate is based, however, on the questionable assumptions that the rate of northwestward thickening of the Snow-*

*bird is approximately known, and that no important erosion of the group occurred before faulting.*"

## Discussion of Hadley and Goldsmith

Observation 1 need not be a definitive criterion in view of the marked lithologic and ductility contrast between the Great Smoky and Snowbird Group, as Hadley and Goldsmith illustrate in their Figure 36, p. B95 (also discussed by King, below). Fold styles in an interlayered sequence of contrasting lithologies (e.g., shale and sandstone) and competency often result in disharmonic folding (e.g., Ramsay and Huber, 1987; Hobbs et al., 1976). The marked discordance in strike of bedding in the Great Smoky Group of the hanging wall in several places (Figure 3, boxes A and B) is admittedly difficult to reconcile without slip along the contact of some magnitude. However, there are areas where bedding in footwall and hanging wall are parallel to the contact, e.g., west of Greenbrier cove and around the Waterville Syncline, and the regional scale pattern is one of general concordancy (Figure 3).

Observation 2 is merely a consistency argument if the contact is a premetamorphic fault. Isograds intersect lithologic contacts in most metamorphic terranes (consistent with the metamorphic facies concept: Turner, 1981), and cleavage commonly diffracts across lithologic contacts. Cleavage in lowest grade rocks in the footwall is actually a weak incipient slaty cleavage defined by chlorite and "sericite" (actually phengite; Clemons and Moecher, in subm. GSAB) and deformed quartz, alkali feldspar, and plagioclase. Hadley and Goldsmith (1963) do not describe in detail the synmetamorphic porphyroblast-matrix microstructural relations for rocks at or above biotite grade, which are mostly in less pelitic, more feldspathic and more competent Great Smoky group lithologies. Our observations indicate that biotite and garnet overgrew an unfoliated to weakly foliated matrix in Great Smoky group metasandstones, and that kyanite overgrew a well-foliated matrix (Clemons and Moecher, in subm. GSAB). Growth of metamorphic index minerals from biotite to kyanite grade occurred before formation of the late 'slip cleavage', as evidenced by deformation of matrix phases around porphyroblasts (Hadley and Goldsmith, 1963; Clemons and Moecher, in subm. GSAB).

Close inspection of an exposure of the contact discovered near the type locality (Clemons, 2006) reveals post-metamorphic ductile deformation of an earlier muscovite-chlorite schistosity in footwall rocks and formation of an S-C fabric or shear band cleavage (Clemons, 2006), which supports post-metamorphic slip. Furthermore, the orientation of cleavage close to (within 100 m) the fault isn't clear from the 1:62,500-scale map of Hadley and Goldsmith (1963). There exists a conspicuous fracture cleavage at the top of the footwall (Clemons, 2006), parallel to a retrograde foliation and not of an axial-planar nature that dips SSE approximately parallel to the fault. Hadley and Goldsmith (1963) apparently interpreted these fabrics as bedding. This foliation is also consistent with late slip along the contact.

With regard to the regional slaty cleavage, Hadley and Goldsmith (p. B75) note that slaty cleavage in the footwall (Pigeon Siltstone and Roaring Fork Formation), even in the area of major folds (Cartertown and Copeland Creek anticlines; Figure 3), is variably developed, i.e., it is not regionally penetrative and even "*inconspicuous*", but "*increases in intensity*" to the south (toward the Greenbrier fault). Cleavage is absent in sandstone beds of Snowbird units. The orientation of slaty cleavage is consistently parallel to strike of more shallowly dipping southern limbs of folds but highly oblique to strike of steeply dipping beds on northern limbs. This is not an axial plane cleavage (Figure 5); the cleavage strikes NE whereas fold axes and bedding-cleavage intersections strike ENE. Rocks of the Elkmont and Thunderhead Fms. above the fault, are coarser grained feldspathic arenites and are not cleaved but weakly foliated (below). Cleavage in interbedded siltstones of the Thunderhead Fm. is accounted for by bedding plane slip in thick metasandstone beds during folding and is unrelated to cleavage in the footwall. In contrast, higher grade rocks to the south in the Cherokee-Dellwood area exhibit schisto-

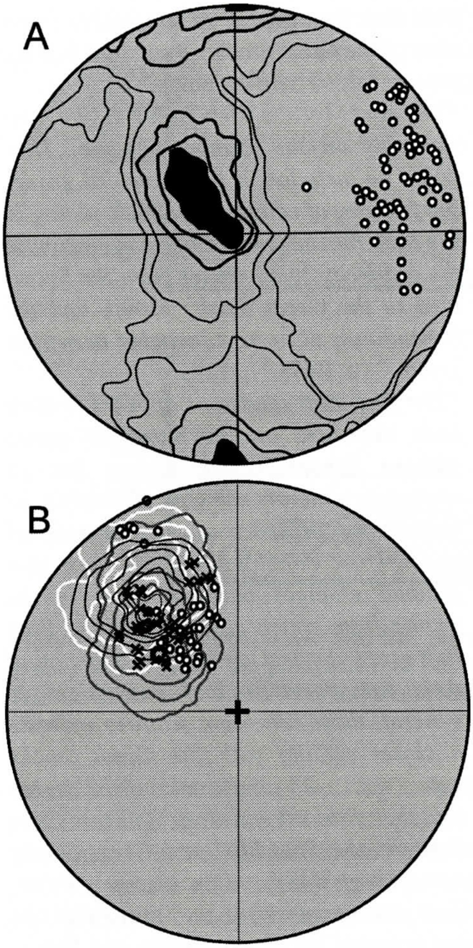



**Figure 5. Orientation of structural elements in the Foothills belt (lower-hemisphere equal area projections). A: Orientation of poles to bedding (contours; Hamilton, 1961) and fold hinge lines (circles; Hadley and Goldsmith, 1963) for cleaved Pigeon Siltstone in the areas of Copeland Creek and Cartertown anticlines (Figure 3); B: Orientation of poles to bedding ($S_0$) (circles: Clemons, 2006); orientation of poles to $S_2$ foliation in Roaring Fork and Pigeon in Greenbrier footwall (crosses: Clemons, 2006); orientation of incipient slaty cleavage in the Pigeon Siltstone in Copeland Creek (white) and Cartertown (gray) contours (2%, 6%, 10% and 20%) (Hadley and Goldsmith, 1963); orientation of incipient slaty cleavage in Pigeon siltstone in Richardson Cove quadrangle (Hamilton, 1961; thin black contours).**

sity and gneissic foliations that strike NE and are axial planar to late folds (Massey and Moecher, 2005).

The minimum 23 km of offset (observation 3) calculated from the inferred stratigraphic thinning of the Ocoee (Figure 10 of Hadley and Goldsmith, 1963; Figure 4 here) is based on a southward thinning of the Longarm Fm. from Big Creek to Cove Creek. This area of greatest thinning corresponds to an area of isoclinal folding and extreme post-metamorphic ductile attenuation of Ocoee units. The area of presumed greatest thinning is beneath the Greenbrier/Caldwell Fork fault defining the southeast margin of the Cataloochee Anticlinorium (Figures 2, 3). Hadley and Goldsmith admit, however, that:

"*...the amount of displacement therefore could be considerably less than the width of its outcrop belt. Nevertheless, the displacement accounts for the great difference in thickness of the Snowbird group on opposite sides of the fault near Cove Creek Gap* [southeast side of Cataloochee Divide synclines; Figure 3] *and is therefore considerable.*" (p. B80)

The estimate of displacement is based on the questionable assumption (quoted in Observation 3 above) of constant thickening and absence of substantial erosion. We include the additional questionable assumption that no significant attenuation of the Snowbird occurred during folding, metamorphism, or subsequent shearing at its southeasternmost outcrops, and that the section is not anomalously thick due to repetition by thrusting or tight folding in the Cataloochee anticlinorium.

Nowhere in their report do Hadley and Goldsmith (1963) describe a specific fault contact, and nowhere do they use the terms ductile or brittle in relation to potential fault rocks at the contact. They reported the mylonites associated with the Cold Spring and Greenbrier faults in the Cataloochee anticlinorium region, but such mylonites are clearly retrograde (Montes, 1997; Drew et al., 2008). Montes (1997) and Montes and Hatcher (1999) observed that the fault mapped as the Greenbrier by Hadley and Goldsmith on the southeast limb of the Cataloochee anticlinorium (Figure 3) is a postmetamorphic fault exhibiting retrograde (greenschist facies) microstructures that offset foliation, the kyanite isograd, and older faults. Hadley and Goldsmith (1963), Southworth et al. (2005b), and Merschat et al. (2006) recognize similar late faults and high strain zones in the Cataloochee area and to the northeast. Montes and Hatcher (1999) renamed this retrograde fault the Caldwell Fork fault, and propose that it is a reactivated Neoproterozoic normal fault northwest of which the thicker Longarm section was deposited. The Greenbrier (or Caldwell Fork) fault merges to the northeast with the retrograde ductile Cold Springs fault (Figure 3), which merges with the Meadow Fork (O'Hara, 1990) and Rector Branch ductile thrust faults south and west of the Hot Springs Window (Merschat et al., 2008). Syntectonic white mica in mylonites of the Cold Springs fault yields $^{40}Ar$-$^{39}Ar$ plateau ages of 330-350 Ma, interpreted to be the time of ductile deformation on the Cold Springs and (by correlation of metamorphic grade and structural relationships) with the Caldwell Fork fault and broad zones of retrograde deformation throughout the Great Smoky Mountains region (e.g., Metcalf phyllite: Kunk et al., 2006; Southworth et al., 2005a). The crucial point of these relationships is that the Greenbrier fault as originally defined in the area of Cataloochee anticlinorium was proposed to be the thrust fault along which there occurred 23 km of premetamorphic displacement of Great Smoky, some Snowbird, and basement over Snowbird group. However, this fault cannot be a putative premetamorphic Greenbrier Fault, which negates one of the key assumptions of the Hadley and Goldsmith model for the geologic evolution of the Great Smoky Mountains region.

Another of the key stratigraphic problems of Great Smoky geology that lacks a definitive solution yet bears on the origin of the contact is the manner in which the Rich Butt Fm. and rocks of Webb Mtn. and Big Ridge fit into Ocoee stratigraphy (Figs. 2, 4). The Rich Butt Fm. crops out mainly around the Waterville syncline at the top of the Greenbrier footwall (Figure 3). Two statements by Hadley and Goldsmith regarding stratigraphic relations of the Rich Butt Fm. permit an alternative interpretation of the Great Smoky-Snowbird contact (underscore added here for emphasis):

1. *"The rocks* [of the Rich Butt Fm.] *lie conformably on the Pigeon siltstone...They could be included in the Snowbird group, but features of composition and bedding in the Rich Butt suggest rather <u>a transition</u> in the sedimentary character from the Snowbird to the Great Smoky group, and the formation is at present included in neither group."* (p. B47)
2. *"The Rich Butt sandstone appears to have been deposited on the Snowbird group <u>without important interruption</u>, but its accessory minerals and type of bedding, as well as the presence of intraformational and arkosic conglomerates, suggest that gradual changes occurred both in the source from which sediments of the Rich Butt were derived and in the manner of their transportation and deposition. In general, these lithologic features indicate a closer affinity with the Great Smoky group than with other rocks of the region and therefore suggest a <u>vertical transition</u> between the Snowbird and Great Smoky groups."* (p. B49)

These statements strongly imply that the contact between the Great Smoky and Snowbird Groups was originally stratigraphic along its entire length and not merely at its southernmost extent where it is shown to be conformal and where there occur exposures of the contact (Figure 4a).

Finally, inconsistencies are noted between Hadley and Goldsmith's (1963) cross sections and their 1:62,500 map (their Plate 1). The map does not everywhere show the orientation of slaty cleavage in the footwall near the inferred fault. The map shows the orientation of bedding with the same orientation as what we observe to be cleavage. Cross sections on King's Plate 1 (B-B', D-D', E-E', G-G', and I-I'; e.g., Figure 6 AA' and BB') show bedding in the footwall as generally parallel to bedding in the hanging wall at the surface, but bedding is then shown in interpreted cross sections as being truncated at depth by the Greenbrier fault. Although we ob-

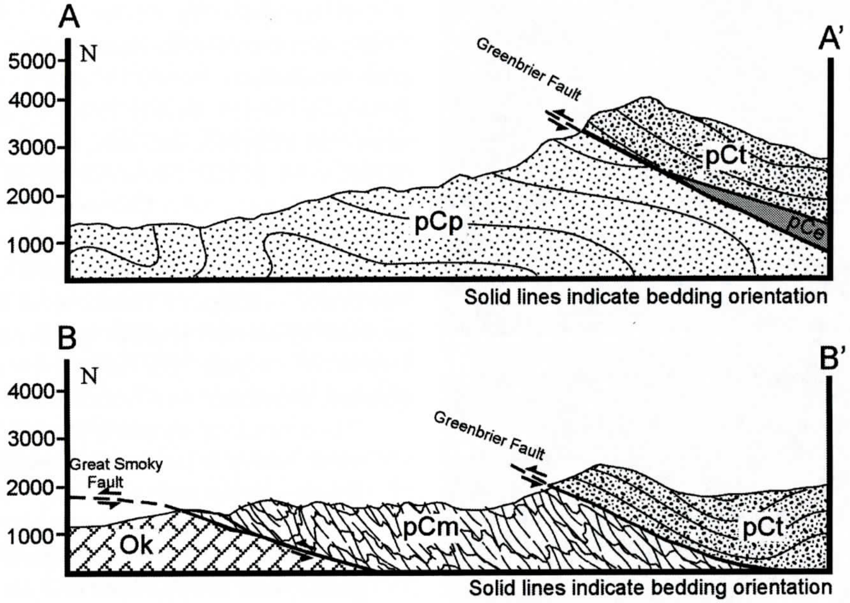



**Figure 6. Cross sections AA' and BB' (locations shown on Fig 2B) form Hadley and Goldsmith (1963). In AA', note the parallelism of bedding in the Pigeon Siltstone (pCp) and the Greenbrier fault at the surface and inferred truncation of the Pigeon Siltstone by the Greenbrier fault at depth.**

served steep truncations of bedding in the footwall near Greenbrier Cove (Clemons and Moecher, in subm. GSAB; Southworth et al., 2005b), we see no evidence from the map to justify their inference of truncation at depth.

## U.S.G.S. Mapping: King (1964) and King et al. (1968)

P.B. King did not explicitly evaluate evidence for the existence of the Greenbrier fault in his Professional Paper 349-C (1964), although the Greenbrier figures prominently in his geologic map of the central Great Smoky Mountains area, and he repeatedly comments on its characteristics. The Greenbrier defines the Great Smoky-Snowbird contact throughout the central map area, including where Thunderhead and Elkmont Fms. are in contact with Metcalf phyllite (Figure 2B). However, several statements and observations by King reveal his uncertainty regarding features related to the Greenbrier fault:

1. "*Exposures within the report area* [central Great Smoky Mountains] *are not sufficient to demonstrate the fault structure, and interpretations must be made on analogy with better demonstrated relations farther east*", [i.e., in the area of report 349-B] (p. C107); and "*...exposures of the fault surface itself are rare. Little information is thus available as to details of deformation related to faulting.*" (p. C112)
2. In his description of the U.S. Hwy. 441 exposure of the Greenbrier fault in downtown Gatlinburg, TN, King comments that the Pigeon siltstone

   "*...has been converted to dark greenish-gray mylonite, which is recrystallized and schistose and contains embedded angular silty fragments, as well as a few larger lenses of overriding Thunderhead...The mylonite is also slickensided, but evident by movements later than the recrystallization.*"

(Figure 7). These are clearly post-metamorphic textures similar to those we observe (Clemons and Moecher, in subm. GSAB) in rocks near Greenbrier Cove, Big Creek, and

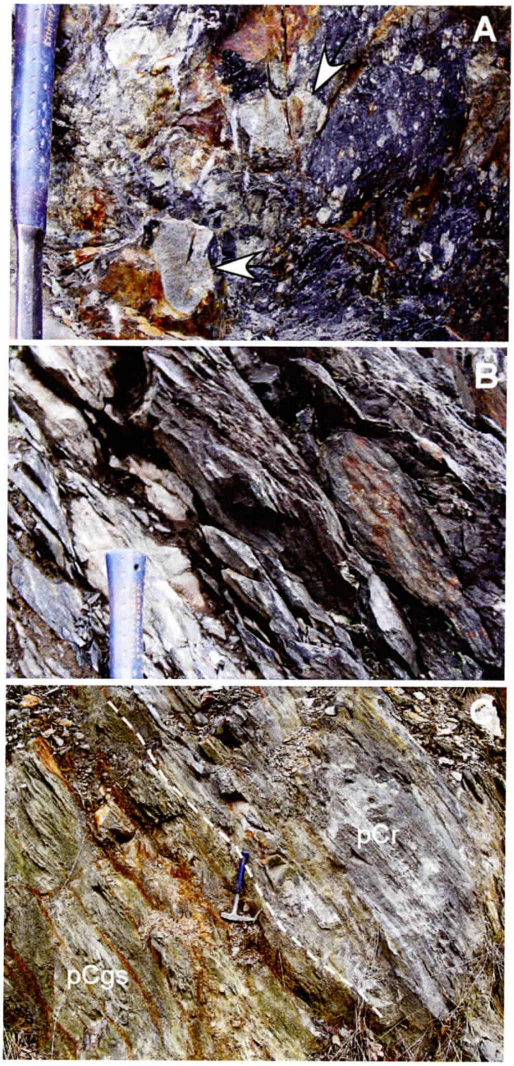



**Figure 7. Field photographs of Great Smoky-Snowbird contact. A. Breccia consisting of clasts of Great Smoky sandstone in sheared, fine-grained phyllonitic Pigeon metasiltstone matrix; Greenbrier Fault exposures in Gatlinburg, TN., originally described by King (1964). B. Shear band cleavage developed in Pigeon Fm. phyllite in footwall approx. 30 m below breccia in A. C. Overturned, conformable Great Smoky (pCgs)-Snowbird (pCr: Roaring Fork) contact near Cove Creek study area (Figure 2A) and section 7 shown in Figure 4.**

Metcalf Bottoms.

3. King recognizes that

   *"The superposition of Thunderhead on Metcalf is certainly tectonic rather than stratigraphic."* (p. C109)

Slip associated with superposition must be post-metamorphic as well because it converts a greenschist facies, weakly cleaved Pigeon siltstone into a phyllitic tectonite, the Metcalf Fm. (King, 1964, p. 109; Southworth et al., 2005b; Woodward et al., 1991; Clemons, 2006).

King describes similar conversion of the Thunderhead formation to tectonic schists at "the Sinks", a klippe of Thunderhead formation enclosed by Metcalf phyllite, and in the "Raven Den slices" (Figure 2B), fault slivers of Thunderhead, imbricated with Metcalf, as follows;

> *"As a result of shearing, the sandstone of these bodies is pervaded throughout by a southeast-dipping foliation which has nearly obliterated the bedding, has shattered and stretched the quartz and feldspar grains, and has flattened and made schistose the argillaceous fragments. Large quartz grains have resisted deformation more than the smaller, and stand out as hard knots in the sheared rock."* (p. C111)

This deformation is most likely the same deformation that reconstituted the Pigeon to produce the Metcalf phyllite, which is postmetamorphic. Based on Pigeon-Metcalf contact relationships in King's map area, the Greenbrier is a postmetamorphic fault of very different character than that implied by Hadley and Goldsmith (1963) at the type section (i.e., premetamorphic).

5. King provides additional insights regarding the nature of Greenbrier fault (p. C121) with the following:

   *"The initial fracture, at least in the forward part of the structure, probably developed along the stratigraphic contact between the two groups, because the Great Smoky strata were highly competent, and the Snowbird group strata somewhat less so."*

And:

> [Relative displacement] "*...between originally contiguous rocks above and below the fault may vary considerably.*"

At some places strata above and below the fault are parallel, but at others varying degrees of truncation are apparent (Figure 3, also discussed above). Furthermore, because of younger-over-older thrusting, the amount of absolute displacement along the fault is uncertain. King notes the actual amount may be lesser or greater than the "*20 mi.*" estimate of Hadley and Goldsmith, although the latter claimed only 15 mi. of displacement (23 km).

Although the base of Great Smoky group rocks is truncated, "*...they were probably emplaced as a little deformed plate.*" This implies there was not significant pre-fault folding in the hanging wall. King further implies that much of the deformation in the footwall and hanging wall was syn-faulting:

> "*Emplacement of Greenbrier thrust sheet was therefore accomplished not only by forward movement of the upper plate, but by shortening of the lower plate by folding, and faulting, either immediately before or during the time when it was overridden. Relative displacement along the fault may thus have increased northwestward from its root zone.*"

6. Regarding the timing of emplacement of the Greenbrier thrust sheet and truncation of folds in the Pigeon, King comments: "*The overridden rocks were deeply truncated by the Greenbrier fault...probably as a result of folding which occurred there, just before or during the Greenbrier thrusting.*" (p. C129)
7. Regarding timing of displacement along the fault, King comments
   "*To the north, cleavage in siltstone is geometrically related to minor folds* [i.e., axial planar, and not related to map scale anticlines such as those shown on Figure 3]. *Farther south, where strata are homoclinal, it dips consistently southward across the nearly vertical beds, but its strike diverges curiously northeastward from that of bedding. The folds in this area may have been formed in rocks overridden by the Greenbrier thrust sheet at about the time of its emplacement; the cleavage superposed on the rocks <u>after they have been folded</u> by differently aligned forces.*"

This is consistent with the orientation of structural elements compiled by Hadley and Goldsmith (1963); intersection lineations produced by bedding and cleavage and fold axes are not parallel to the axial plane cleavage (Figure 6).

Additional statements by King et al. (1968), listed below, more explicitly indicate King's interpretation of the nature of the Great Smoky-Snowbird contact:

1. "*...a definitive reconstruction cannot yet be made of either the original relations of the groups or their history...some speculations can be offered*", based on the "*facts*" that: (1) to the southeast the Snowbird group is thin and is overlain conformably by Great Smoky group; and (2) to the northwest the Snowbird group is thick and is overlain conformably by unclassified formations of Ocoee Series (i.e., Rich Butt Fm.), which may be a vertical and lateral transition from the Great Smoky to the Snowbird Group.
2. The Greenbrier fault lies along the Great Smoky-Snowbird contact, which "*...was their original stratigraphic arrangement.*" (p. 12)
3. "*The actual surface of the fault is exposed in few places, but its existence is indicated by contrasting structures in rocks above and below it, and by truncation of these structures at the fault...*" (p. 12)

Based on these observations and comments, it is reasonable to conclude that King interpreted the Greenbrier as a faulted stratigraphic contact along which an unconstrainable amount of displacement occurred.

## U.S.G.S. Mapping: Hamilton (1961)

Hamilton (1961) mapped the Richardson Cove and Jones Cove quadrangles, immediately north of the western end of Hadley and Goldsmith's map area. The southern half of Hamilton's map area is underlain mostly by folded, faulted, and weakly cleaved Pigeon siltstone and the rocks of Webb Mtn. and Big Ridge, whose stratigraphic affinity was and remains speculative. The map area also includes the type localities of most of the formations comprising the Walden Creek Group, and Paleozoic strata deformed by late Paleozoic folding and cleaving. The Greenbrier fault lies to the south of the map area. However, the petrofabrics and mineral assemblages of the Pigeon siltstone in these quadrangles serve as the basis for comparison of the pre- and post-metamorphic history of the Greenbrier fault in the Great Smoky Highlands. Evidence of metamorphism in these sub-greenschist to greenschist facies rocks is the presence of weak, incipient slaty cleavage defined by parallelism of neoformed phengite and chlorite (Clemons and Moecher, in subm. GSAB).

For rocks south of the Dunn Creek fault (mostly Pigeon siltstone), equal area stereonet diagrams (Figure 6) show that cleavage strikes generally NE in both Richardson and Jones Cove quads, with the highest concentration striking 045 and dipping 45 SE, and that the best fit fold axis to the girdle - distributed poles to bedding trends 080. Most major folds in the Pigeon in the southern half of Hamilton's report area are therefore considered premetamorphic; slaty cleavage was superimposed on already deformed rocks. In this regard, Hamilton states that

*"...the divergence between strikes of cleavage and axial planes of these folds is obvious..."*

Hamilton further articulates the patterns evident in the structural data of Hadley and Goldsmith (Figure 6) with the following:

"*Divergence in strike of cleavage and axial planes shows that folds* [in the Pigeon] *and cleavage could not have developed simultaneously. Cleavage is thus a feature superimposed on the folds of the south half of the area. The cleavage developed from pervasive microshearing and mineralogical reconstitution, essentially without further folding, superimposed on a differently oriented system of* [northward] *overturned folds.*"

This diachroneity was also noted by Hadley and Goldsmith (1963) and King (1964) and begs the question of the origin of the slaty cleavage and how it relates to the metamorphic-deformation history. This question is addressed in Clemons and Moecher (in subm. GSAB).

The key aspect of Hamilton's work relevant to the nature of the Greenbrier fault is the relationship between rocks of Webb Mountain and Big Ridge (WM/BR) and the Pigeon, Rich Butt, Elkmont, and Thunderhead formations. Although bedding is vertical and the northern contact of WM/BR rocks and the Pigeon Fm. is a thrust fault, Hamilton seems confident that WM/BR rocks were originally conformable with the Pigeon, based on parallelism of bedding along the south flank of Webb Mtn. However, conflicting interpretations of stratigraphic correlations are expressed (p. A-18). Based on a similarity of lithology and structural position of the lower division WM/BR rocks with the Rich Butt Fm.; Hamilton states

"Correlation of the Rich Butt with the lower division of the Webb Mountain-Big Ridge rocks thus seems likely."

The faulted base of Webb Mountain cannot be correlated with the Greenbrier fault, as is done by other workers (e.g., Woodward et al., 1991). However, based on the lithologic similarity of lower and upper division WM/BR rocks with Elkmont and Thunderhead formations, Hamilton also states the correlation of the two pairs is "probable". This conclusion contradicts Hamilton's interpretation that the WM/BR-Pigeon contact is conformable, and permits the interpretation that the faulted base of Webb Mountain and Big Ridge is an extension of the Greenbrier fault.

## Subsequent Research

Woodward and others (1991; Walters, 1988; Connelly, 1993; Connelly and Woodward, 1992; Connelly and Dallmeyer, 1993), interpret thrust faults (Greenbrier, Dunn Creek and Miller Cove thrusts) in Ocoee stratigraphy as a Taconian foreland ramp-flat system. Restoration of inferred thrust sheets from youngest to oldest led to the inference that the Snowbird and Great Smoky groups are lateral, chronostratigraphic equivalents deposited in an Ocoee basin of apparently much wider extent than the present outcrop width. This interpretation is based on assumed correlation of rocks on Webb Mountain and Big Ridge with the Great Smoky group, and the Cades formation to the west with the Elkmont formation of the Great Smoky group. Webb Mountain and Big Ridge are interpreted to be klippe of Great Smoky group equivalents, based on truncations of bedding and fault-parallel mylonites at the base of the hanging wall (Connelly et al., 1989). The inferred correlations and restoration of Great Smoky group-equivalents back to the "*Snowbird-Great Smoky facies change*" permits a minimum displacement of 23 km for the Greenbrier thrust sheet, similar in magnitude to the estimate of Hadley and Goldsmith (1963) based on the purportedly missing Longarm Fm. section in the eastern GSM.

Woodward and others recognized that cleavage diverges from axial planes of folds throughout the Foothills belt and was superimposed on early folds, as recognized by Hamilton (1961), Hadley and Goldsmith (1963), and King (1964). Connelly et al. (1989) established that this cleavage is axial planar to folds in the Miller Cove thrust sheet, but transects most other structures within the Dunn Creek and Greenbrier thrust sheets. Woodward et al. (1991) correlate regional metamorphism with formation of cleavage through the study area. Based on this assumption, the cleavage exhibited by rocks in the Miller Cove thrust sheet was inferred to have been superimposed on folded but non-penetratively deformed rocks of the Greenbrier hanging wall and that the Miller Cove fault was active at the time of metamorphism.

Several other observations warrant noting. (1) Woodward et al. (1991), and Southworth et al. (2005b) recognize that the Metcalf phyllite is a tectonite. (2) Post-metamorphic mylonites and shearing are recognized, particularly at the base of the Webb Mtn. thrust block (Connelly, 1989) and bounding the package of Cades Fm. in contact with Metcalf phyllite (Walters, 1988). These must be post-metamorphic structures, and are similar to late features described by King (1964), Southworth et al. (2005b), and Clemons and Moecher (in subm. GSAB).

Although there are many aspects of this work with which we agree, there exist shortcomings in assumptions on which the restoration approach is based. We disagree with the following points:

1. All structures are interpreted to be related to foreland fold and thrust belt style deformation that occurred before the peak of metamorphism. The deformation and metamorphic history of higher grade Snowbird and Great Smoky units in the southern Great Smoky Mountains is not incorporated into the model. It is implied that there is only minor post-Taconian overprinting and Woodward et al. assume that all regional cleavages (even into the kyanite zone) are of a single generation. This is clearly not the case. Based on observations presented elsewhere, (Massey and Moecher, 2005; Moecher et al., 2005; Clemons, 2006; Clemons and Moecher, in subm. GSAB), there exists: (1) sub-biotite-grade, incipient slaty cleavage spatially associated with (but not genetically related to) folds in the footwall (Pigeon Siltstone) whose relation to regional metamorphism is uncertain due to lack of index minerals; (2) chlorite-grade post-metamorphic (biotite grade) slaty cleavage in Roaring Fork, Pigeon, and Rich Butt Fms., in the footwall of the Greenbrier; (3) cleavage within thin shale interbeds in the Thunderhead Fm. related to bedding plane slip during late folding; (4) schistosity related to prograde metamorphism, and retrograde spaced 'slip cleavage' (Hadley and Goldsmith, 1963) that develops southward into a fold-related crenulation cleavage (Massey and Moecher, 2005; Moecher et al., 2005). The rocks south of the garnet isograd were isoclinally folded and ductilely sheared during retrograde metamorphism, and did not behave as a rigid thrust sheet, although the Great Smoky rocks were certainly more competent than Snowbird rocks. Therefore, the approach of Woodward et al. may apply only to the area north of the Alum Cave Syncline (Figure 3) if incipient slaty cleavage were an axial plane foliation, but this is not the case. South of this area the rocks are higher grade garnet to kyanite schists overprinted by a subsequent foliation (Clemons and Moecher, in subm. GSAB).
2. The use of cleavage as a temporal refer-

ence for pre- or postmetamorphism is unwarranted as biotite, garnet, staurolite and kyanite porphyroblasts exist throughout the study area and nearly all foliations deform or wrap around porphyroblasts. Porphyroblast growth should be the reference for pre- versus postmetamorphic timing.

3. If the Great Smoky group-Snowbird group lateral chronostratigraphic equivalency is valid, one would expect to observe lateral intertongueing of the two groups. Lateral and vertical facies variations are apparent *within* each group (Elkmont-Thunderhead-Anakeesta; Roaring Fork- Pigeon-Rich Butt: Hadley and Goldsmith 1963; King, 1964) (Figure 4), but not *between* the groups. Descriptions of the Rich Butt Fm. suggest a vertical gradation (see above).
4. Woodward et al. accept the 420-460 Ma $^{40}Ar$-$^{39}Ar$ whole rock slate and phyllite ages of Connelly and Dallmeyer (1993) as the time of metamorphism for the northern Great Smoky Mountains. The ages have no geologic meaning because they are whole rock ages from low grade rocks containing multiple sources of Ar that has not been completely purged by prograde metamorphic recrystallization and reaction. Woodward et al. assume that all deformation is either Taconian or Alleghanian, and no intervening Acadian or Neo-Acadian effects are considered. Although peak metamorphism was Taconian (Moecher et al., 2004, 2005; Kunk et al., 2006; Anderson and Moecher, 2008; Corrie and Kohn, 2007), and overprinting is indeed relatively weak in the Foothills belt up through garnet grade (Clemons and Moecher, in subm. GSAB), postmetamorphic overprinting dominates the deformation history of rocks above garnet grade and becomes extremely intense above staurolite grade (Massey and Moecher, 2005; Clemons and Moecher, in subm. GSAB). "Neo-Acadian" (late Devonian to early Carboniferous) tectonism is becoming increasingly recognized in the region (Merschat et al., 2006; Hatcher and Merschat, 2006), and is the likely time of retrograde deformation (Moecher et al., 2005, 2007; Kunk et al., 2006).

## Recent U.S.G.S. Investigations

The U.S. Geological Survey re-examined the surficial and bedrock geology of Great Smoky Mountains National Park, and recompiled existing 1:24,000 scale map coverage into a revised and internally consistent 1:100,000 scale map (Southworth et al., 2005b). Key observations and conclusions relevant to this paper include:

1. The Rich Butt Formation, previously an "unassigned" stratigraphic unit is assigned to the upper Snowbird Group. Its primary clastic components, accessory minerals, and bedforms are consistent with the Rich Butt being a vertically transitional unit between the Snowbird and Great Smoky groups, as considered by Hadley and Goldsmith (1963, see above). Southworth et al. (2005b) also correlate other "unassigned" units on Webb Mtn. and Big Ridge to the Rich Butt, and not the Great Smoky group as proposed by Woodward et al. (1991). In this interpretation fault rocks at the bases of Webb Mountain and Big Ridge did not result from displacement along the Greenbrier fault.
2. The Metcalf phyllite is the phyllonitized equivalent of the Pigeon siltstone (Figure 8) with structures indicating top to the northwest shear sense. This post-metamorphic deformation occurred within the Little River Shear Zone, which is bounded below by the Great Smoky fault (base) and above by the Greenbrier fault (roof). $^{40}Ar$-$^{39}Ar$ geochronologic analysis of Metcalfe phyllite indicates phyllonitization of Pigeon Siltstone to form the Metcalf Phyllite occurred at approximately 350 Ma (Kunk et al., 2006).
3. Post-metamorphic deformation is widespread in the Great Smoky group and, other than those structures recognized by Hadley and Goldsmith (1963), is manifested in ductile high strain zones such as

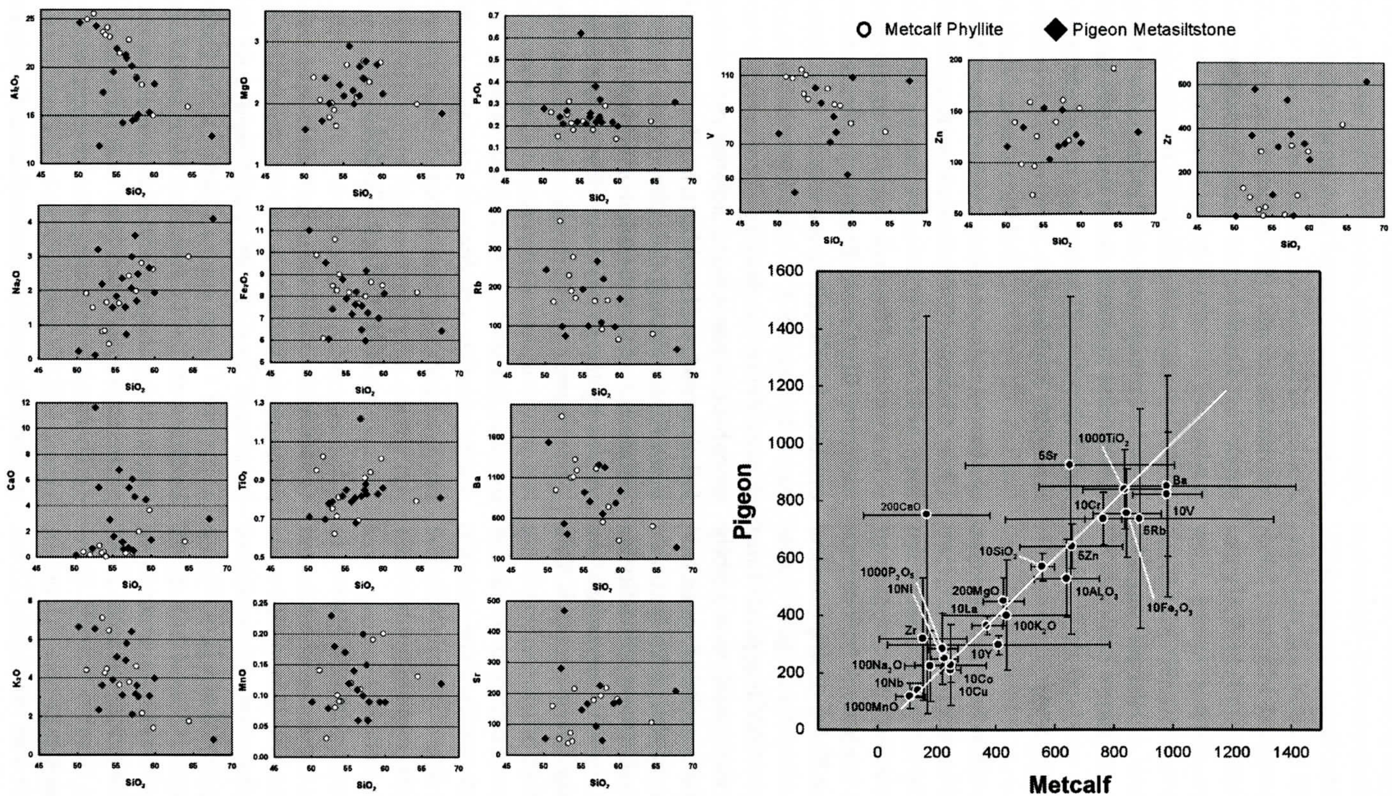



Figure 8. Results of whole rock major, minor and trace element analysis of Pigeon siltstone and Metcalf phyllite (data in Clemons, 2006; sample localities shown in Figure 2B) supporting field and petrographic evidence that the Metcalf phyllite is the tectonized equivalent of the Pigeon siltstone (King, 1964; Southworth et al., 2005; Clemons, 2006). Analyses are of whole rock powders not pretreated to remove carbonate. Scatter plots show general consistency in range of concentrations and degree of heterogeneity for each element. Correlation diagram shows that most elements exhibit the same concentration, on average and within error, in the Metcalf and Pigeon, except CaO and Sr. Some Metcalf samples contain calcite that formed during greenschist facies shearing and recrystallization, which accounts for anomalously high CaO and Sr relative to average Pigeon siltstone samples.

the Eagle Creek and Little River Shear Zones, the Oconaluftee, Gatlinburg, Mingus, Cold Spring, and Caldwell Fork faults, and narrower, unnamed high strain zones in the eastern Great Smoky Mountains (Montes, 1997; Cattanach and Merschat, 2005; Massey and Moecher, 2005; Merschat et al., 2006). $^{40}Ar$-$^{39}Ar$ geochronology on phyllonites in the latter zones yield ages of 350-330 Ma, interpreted as the time of ductile thrusting (Southworth et al., 2005a).

## DISCUSSION

Specific exposures of the fault near the type locality at Greenbrier Pinnacle, where it is least overprinted by regional metamorphism or not reactivated by younger deformation, were never described by previous workers. A fault surface along which more than 20 km of displacement has occurred, with a fault surface highly discordant to bedding (i.e., not a bedding plane thrust along an incompetent shale unit), would likely be characterized by fault rocks of some kind and/or brittle fracturing of wall rocks. No brittle fault fabrics are observed in the type locality for the Greenbrier fault or any other exposure of the Snowbird-Great Smoky Groups contact. This fact and the preservation of relatively delicate structures such as planar and ripple lamination in footwall rocks near the contact, and lack of a strong fault fabric in Great Smoky rocks just above the contact, are inconsistent with a premetamorphic brittle fault.

The only demonstrable relationship cited in previous studies that is consistent with a fault at the Great Smoky-Snowbird Group contact is the variable degree of discordance in bedding of Great Smoky Group and Snowbird Group rocks. This relationship is however only apparent at low grade near the northern contact (we also observed this discordance: Clemons, 2006; Clemons and Moecher, in subm. GSAB). However, the contact is not everywhere discordant (Figure 3). Discordance does not require 23 km of offset and may be explained by an alternative mechanism (discussed below). Furthermore, the proposal that the fault transports thick, younger Great Smoky over older Snowbird group rocks for a distance of approximately 20 km is used to infer that the Ocoee requires two depositional basins, which leads to further speculation on the nature of those basins (Rast and Kohles, 1986) and additional unnecessary complexities in the geologic history of the region. A single basin with lateral facies variations and gradational contacts is a simpler model.

Hadley and Goldsmith (1963) cite no specific outcrop exposure of the fault, but King (1964) cites locations of the fault in the central map area where it is clearly a retrograde structure. The fault is also exposed in accessible roadcuts in the Metcalf Bottoms area, where it is also retrograde (King, 1964; King et al., 1968). Clemons (2006) located exposures of the fault near Greenbrier Cove, and exposures within meters of the fault along Greenbrier Pinnacle and Big Creek. These exposures exhibit evidence for slip along the contact (e.g., shear band cleavage, S-C fabric, S-tectonites) and thus evidence for faulting. However, the required slip is late in that it deforms pre-existing foliations and porphyroblasts, similar to the pattern of overprinting that affected the Pigeon Siltstone to form the Metcalfe phyllite described by King (1964; also Southworth et al., 2005; Figure 8). In summary, regional and contact relations do not require a major premetamorphic thrust fault and an alternative model, discussed below, can account for almost all the observed geologic relationships.

### Alternative Model

The minimum 23 km of premetamorphic offset proposed by Hadley and Goldsmith (1963) is inferred from the thickness change of the Snowbird Group across a fault in the Cataloochee and Cove Creek Gap areas originally mapped by Hadley and Goldsmith as the premetamorphic "Greenbrier fault" (Figures 3, 4). This fault is now recognized to occur within a broad NE-SW trending zone of high strain and mylonitization that extends at least as far southwest as Cherokee (Southworth et al., 2005b) and into the Fines Creek and Lemons Gap quadrangles to the east and northeast (Cattanach et al., 2005; Merschat et al., 2008). Although it separates Snowbird and Great Smoky group rocks, it is clearly a ductile postmetamorphic high strain zone (Montes, 1997; Drew et al., 2008), and cannot be a putative premetamorphic Greenbrier fault (Caldwell Fork fault: Montes and Hatcher, 1999). Thus, explanations for the history of the Snowbird-Great Smoky contact must account solely for the local angular discordance of bedding that is evident along the mountain front between Greenbrier Pinnacle and Big Creek to Mt. Sterling, and inferred from mapping in the area of Cove Creek (Montes and Hatcher, 1999). A model for the origin and deformation of rocks along the contact must be couched within a comprehensive model for the postmetamorphic history of the entire Blue Ridge and adjacent terranes. We agree with King (1964), who considers the contact to be originally stratigraphic and to have experienced post-metamorphic displacement of unconstrained magnitude.

Based on existing work and new observations presented by Clemons and Moecher (in subm. GSAB), we propose the alternative hypothesis that the angular discordance observed along Greenbrier Pinnacle resulted from differential folding of footwall and hanging wall lithologies of markedly differing competency, accompanied by slip along the contact to accommodate the differential folding (King, 1964). The hanging wall is competent Great Smoky units in which bedding simply warped. Slip resulted in post-metamorphic ductile deformation in a meter-scale zone at the type locality, and in a much wider, kilometer-scale zone west of Gatlinburg (the Little River shear zone of Southworth et al., 2005b) developed in the Metcalf phyllite and fault slivers and klippe of the Great Smoky Group.

Hadley and Goldsmith (1963) recognized that a second period of folding produced a second generation of folds and schistosity in rocks above approximately garnet grade south of the Alum Cave Syncline (Figure 3). This period of folding generated structures ranging from the major map-scale folds in the Bryson-Cherokee-Dellwood-Sylva area (Figure 3) to the thin section-scale 'slip cleavage' present in the Pigeon, Roaring Fork, Elkmont, and Thunderhead formations (Clemons, 2006), which grades into a post-metamorphic schistosity in Great Smoky equivalents to the south (Massey and Moecher, 2005; Moecher et al., 2005). Numerous post-metamorphic ductile faults and high strain zones formed at the time of folding. Late shortening was accommodated by folding of footwall, hanging wall, and basement rocks, shortening along reverse high strain zones and folding of the contact into a broad synform (the 'Waterville syncline'; Southworth et al., 2005b). This yields the apparent thrust relationship along the northern mountain front and caused northward extrusion of the great mass of competent Great Smoky group lithologies relative to Snowbird group lithologies. The vertical orientation of metamorphic isograds indicated by map patterns (Figure 3) suggests isograds were rotated upward and northward to shallower levels during north-vergent folding of the higher grade rocks in the Cherokee-Dellwood belt. The age of deformation responsible for folding and uplift is uncertain, but $^{40}$Ar-$^{39}$Ar plateau ages of hornblende and muscovite from staurolite and kyanite grade rocks (450-420 Ma and 370-350 Ma: Dallmeyer, 1975; Connelly and Dallmeyer, 1993; Kunk et al., 2006), and $^{40}$Ar-$^{39}$Ar ages on Metcalf phyllite and retrograde greenschist facies mylonites (350 to 330 Ma; Kunk et al., 2006) suggest uplift and cooling through the closure temperature for Ar diffusion (~350-375 °C: McDougall and Harrison, 1988) is definitely post-Taconian, but could have occurred from Neo-Acadian (late Devonian to early Mississippian) to early Alleghanian time (late Mississippian to early Pennsylvanian). The late deformation may be correlated with and driven by transpressional tectonism of a similar age in the adjoining Eastern Blue Ridge and Inner Piedmont (Hatcher, 2002; Merschat et al., 2005; Hatcher and Merschat, 2006; Moecher et al., 2007).

## SUMMARY

The key assumptions explicated to support a premetamorphic, large displacement ($\geq$ 23 km), younger-over-older Greenbrier fault are largely unfounded. These assumptions include (Hadley and Goldsmith, 1963):

A. Truncation of bedding in the hanging wall and footwall is observed along the northern extent of the fault. Truncation requires displacement of some magnitude, but it is clearly postmetamorphic at all exposures of the fault (King, 1964; Clemons, 2006).
B. The putative root zone of the Greenbrier fault as originally proposed is a retrograde ductile high strain zone related to regional Carboniferous (early Alleghanian) thrusting.
C. The Snowbird-Great Smoky Group is a fault, however statements by original workers clearly state that the prefaulting contact was gradational from Snowbird (Pigeon Fm.) to unclassified units (Rich Butt Fm.) to Great Smoky units. Outcrops of the contact in the Cove Creek area demonstrate that the contact is conformable and gradational, as shown by Hadley and Goldsmith (1963, their Figure 10).

Almost all stratigraphic, structural, and metamorphic features are consistent with a Greenbrier Fault involving postmetamorphic slip of uncertain magnitude (but not tens of km) along a faulted stratigraphic contact between two units of contrasting competency. This model is simpler and does not require multiple Ocoee depositional basins or a Taconian foreland ramp-flat system to account for stratigraphic and structural relationships in the Great Smoky Mountains region.

## ACKNOWLEDGMENTS

We appreciate the insight provided by Scott Southworth (U.S.G.S.), and Carl Merschat and Bart Cattanach (North Carolina Geological Survey).

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SOUTHEASTERN GEOLOGY
V. 45, No.4, August 2008, p. 225-232

# ADDITIONAL EDIACARAN BODY FOSSILS OF SOUTH-CENTRAL NORTH CAROLINA

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## ABSTRACT

**The Carolina Terrane has yielded very few Ediacaran-age fossils. These included *Pteridinium carolinaensis,* cf. ?*Swartpuntia* sp. and a variety of trace fossils. Recent collecting efforts in the Albemarle Group have yielded *Sekwia excentrica* from the Floyd Church Formation, an indeterminate sac-like organism from the unnamed Mudstone Member of the Cid Formation, Stanly County and *Aspidella* cf. *A. terranovica* from the unnamed Mudstone Member of the Cid Formation, Davidson County, North Carolina. North Carolina Ediacara biota, though considered deep-water, has greatest affinity with the Nama assemblage of similar age.**

## INTRODUCTION

Ediacaran-age fossils are known from approximately thirty localities on five different continents (Narbonne, 2005). A number of studies have focused on what their distributions patterns reveal about paleogeographic reconstructions (McMenamin, 1982; Donovan, 1987; Waggoner, 1999, 2003; Hagadorn and Waggoner, 2000), paleoenvironments, paleoecology, and taphonomy (Grazhdankin and Ivantsov, 1996; Gehling, 2001; Clapham and Narbonne, 2002; Waggoner, 2003; Grazhdankin, 2004, Narbonne, 2005).

Recent collecting efforts in the Albemarle Group of Carolina Terrane by Ruffin Tucker, Tony C. Furr, researchers from the North Carolina Museum of Natural Sciences and North Carolina State University have yielded additional specimens of Ediacaran body fossils, including the probable Ediacara holdfast, *Sekwia excentrica* Hofmann, 1981, a possible sac-like organism, and *Aspidella* cf. *A. terranovica* from Neoproterozoic rocks of North Carolina. Though *Asipidella* cf. *A. terranovica* was mentioned by Hibbard et al (2006) and Weaver et al (2006b) it was never formally described. These new discoveries combined with previously described cf. ?*Swartpuntia* sp. by Weaver et al. (2006a), *Pteridinium carolinaensis* by St. Jean (1973); Gibson et al. (1984); Gibson and Teeter (2001); and McMenamin and Weaver (2002) and trace fossils such as *Planolites* described by Gibson (1989) give a more complete picture of the North Carolina Ediacara biota. The Ediacara fossils from the Carolina Terrane are considered a Nama Assemblage because of similarities in age, the presence of *Pteridinium* (St. Jean, 1973; Gibson et al., 1984; Gibson and Teeter, 2001; McMenamin and Weaver, 2002) and the presence of true trace fossils (Gibson, 1989). Thus, though nowhere near as extensive or complete as many of the other Ediacaran Assemblages world-wide, the Carolina Terrane fossils are important as they are a record of a deep-water (Seilacher et al., 2005) Nama As-

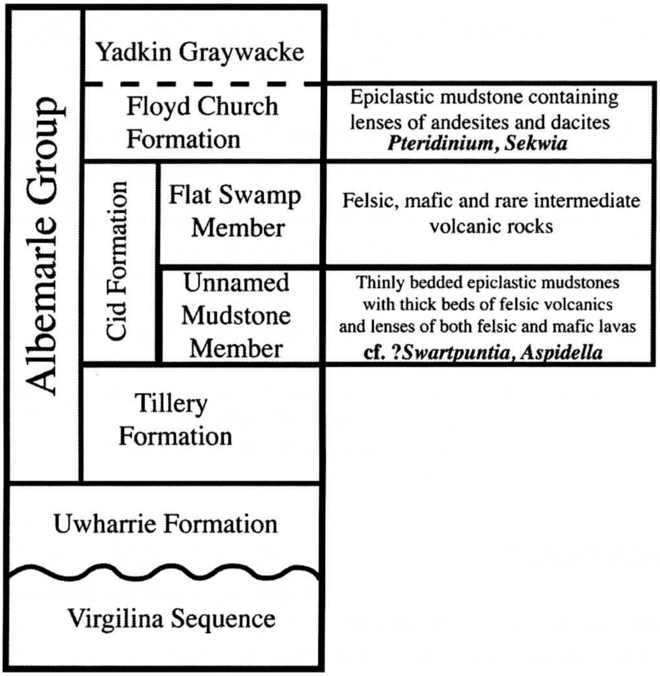



**Figure 1. Generalized stratigraphic section of the Carolina Terrane exposed in this study area.**

semblage.

## GEOLOGIC SETTING

The Carolina Terrane extends from southern Virginia southwest to Georgia and is considered to be one of several peri-Gondwanan exotic terranes that accreted to the eastern margin of Laurentia during the Paleozoic (Hibbard et al., 2002). Rocks within the Carolina Terrane show a lower green-schist facies grade of regional metamorphism (Gibson and Huntsman, 1988). Stratigraphic reconstructions of the Carolina Terrane in North Carolina have undergone a series of revisions and in North Carolina workers have recognized three main stratigraphic units: the Virgilina Sequence, the Uwharrie Formation and the Albemarle Group (Conley, 1962; Conley and Bain, 1965; Stromquist and Sundelius, 1969; Seiders, 1978; Milton, 1984) (Figure 1). The Albemarle Group has been further subdivided into Tillery Formation, Cid Formation (consisting of the unnamed Mudstone Member and the overlying Flat Swamp Member), Floyd Church Formation and Yadkin Formation (Milton, 1984) from oldest to youngest. Seilacher et al. (2005) considers the Albemarle Group to have been deposited in deep-water as the formations within the group are characterized by thick sequences of laminated mudstones and coarser-grained turbidites, with an absence of wave induced structures.

Neoproterozoic Ediacaran body fossils have been reported from the Floyd Church Formation (St. Jean, 1973; Gibson et al., 1984; Gibson and Teeter, 2001; McMenamin and Weaver, 2002) and from the unnamed Mudstone Member of the Cid Formation (Hibbard et al., 2006; Weaver et al., 2006a). Neoproterozoic trace fossils have been reported from most of the Albemarle Group except the volcanic Flat Swamp Member of the Cid Formation (Gibson, 1989). The new specimen of *Sekwia excentrica* Hofmann, 1981 was found lying on the ground on property surrounding Gleaning Mission Church, Stanly County, North Carolina, by Ruffin Tucker. Based on rock type and locality this specimen most likely came from the Floyd Church Formation (Figure 2). *Pteridinium carolinaensis* has also been reported from this lo-

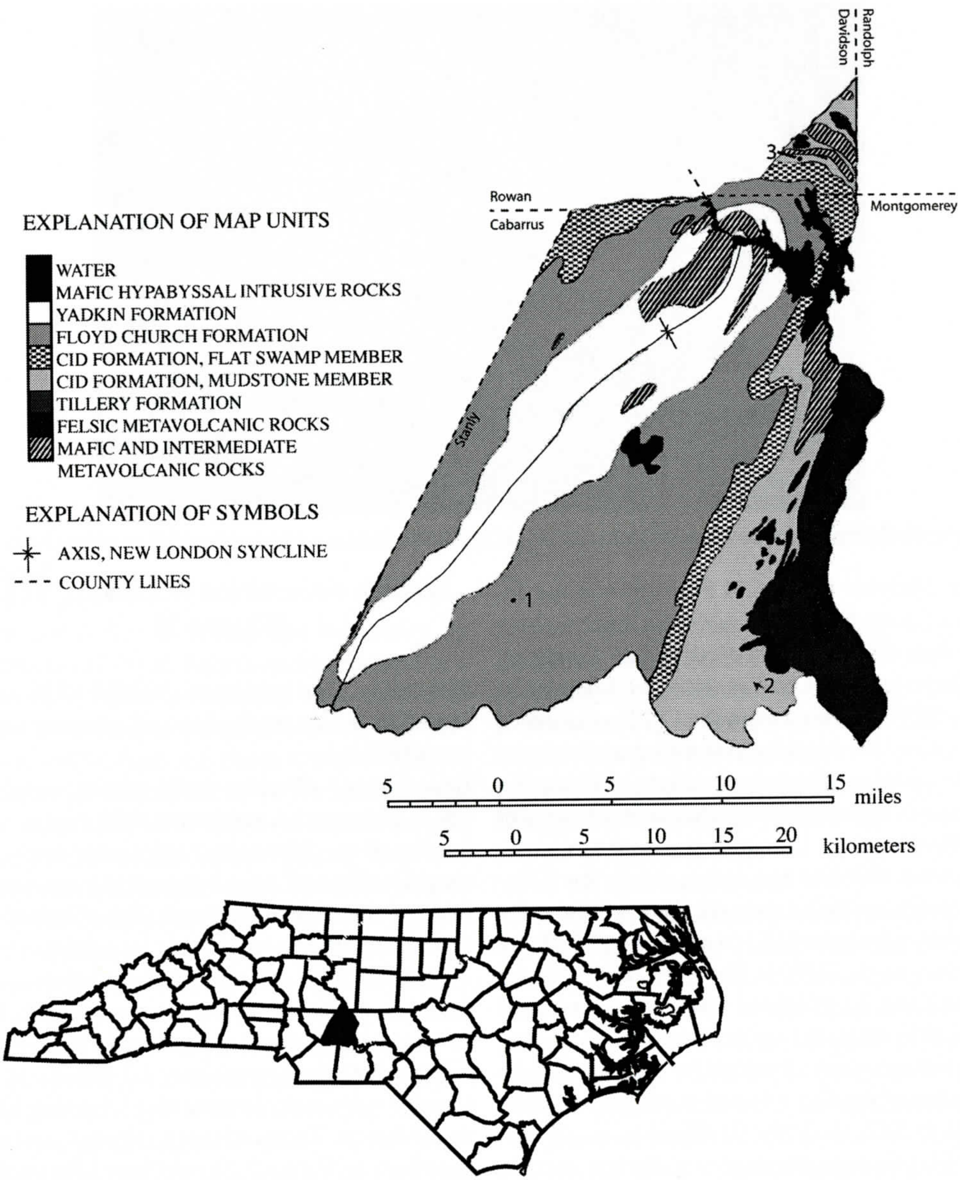



**Figure 2. Map of North Carolina with Stanly County highlighted along with a geologic map of south-central North Carolina (adapted from Goldmith et al., 1988) showing Ediacaran localities (Gibson and Teeter, 2001; Hibbard et al., 2006; Weaver et al., 2006a): 1)Gleaning Mission Church, north of Oakboro, Stanly County, 2) Private property near Mount Zion Church, southeast of Cottonville, Stanly County, 3) Jacobs Creek Stone Quarry, Davidson County.**

cale (Gibson and Teeter, 2001; McMenamin and Weaver, 2002). The new specimen of a sac-like organism was found on private property southeast of Cottonville (near Mt. Zion Church), by Tony C. Furr in the same general area from which cf. ?*Swartpuntia* sp. was reported by Weaver et al. (2006a). Based on rock type and locality the fossil of this sac-like organism most likely came from the unnamed Mudstone Member of the Cid Formation (Figure 2). *Aspidella* sp. cf. *A. terranovica* Billings, 1872 was recovered from the unnamed Mud-

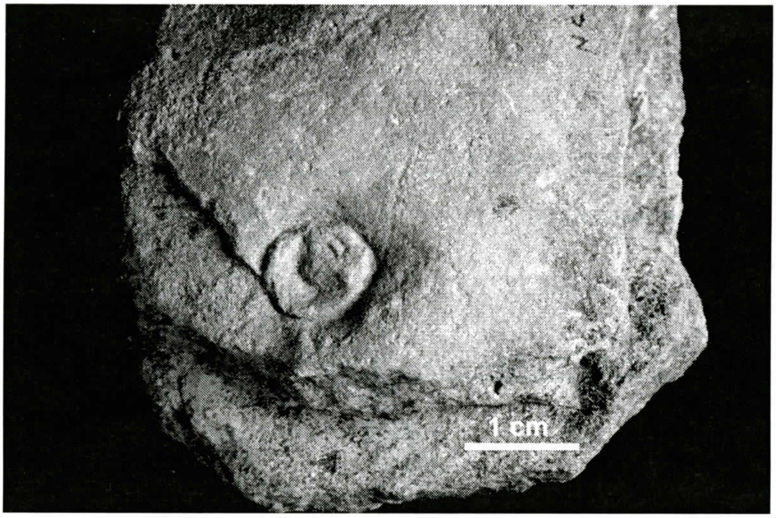



**Figure 3. *Sekwia excentrica* Hofmann, 1981, NCSM 9836 recovered from locality 1 of figure 2.**

stone Member of the Cid Formation at the Jacob's Creek Quarry, Davidson County North Carolina (Figure 2). Reports of late Cambrian or Ordovician age euconodonts by Koeppen et al. (1995) recovered from the Cid Formation of this quarry are considered as either erroneous or representative of structural inliers of Paleozoic sediments on top of Neoproterozoic (Hibbard et al., 2006).

Recent absolute age dating shows the Albemarle Group to be near the Neoproterozoic-Cambrian boundary (given as 542 +/- 0.3 Ma by Amthor et al., 2003). The ages of the Floyd Church and the unnamed Mudstone Member of the Cid Formations are best constrained by recent isotopic dates obtained for the Flat Swamp Member of the Cid Formation, which has been dated at 547 +/- 2 Ma (Hibbard et al., 2006). The Flat Swamp Member is a distinct marker unit which lies stratigraphically between the overlying Floyd Church Formation and the underlying unnamed Mudstone Member of the Cid Formation.

## SYSTEMATIC PALEONTOLOGY

All figured specimens are housed at the North Carolina Museum of Natural Sciences (NCSM) in Raleigh, North Carolina.

### *Sekwia excentrica* Hofmann, 1981 (Figure 3)

*Material*— One specimen, NCSM 9836, preserved as a cast on the base of a brown sandstone bed.

*Description*— Circular fossil approximately 1 cm in diameter. A crescentic fold cuts across the fossil to delineate a relatively smooth, roughly elliptical area between the crescentic fold and the edge of the fossil. The elliptical area has a dimple or crease near its center.

Viewed from stratigraphically below (i.e., from the underside of the sandstone bed), the fossil sits in a 2 mm depression and has a maximum relief of approximately 1.3 mm.

*Locality*— Property surrounding Gleaning Mission Church, Stanly County, North Carolina, Locality 2 of Figure 2, Floyd Church Formation

*Remarks*— *Sekwia excentrica* Hofmann, 1981, which Hoffmann originally presumed to belong with the coelenterates, most likely represents the remains of a holdfast similar to *Aspidella*. The crescentic fold that is diagnostic for the genus most likely resulted from the stalk being tugged, probably by current drag applied to the main body and stalk of the organism.

Although the morphology of this Ediacaran fossil appears to be quite simple, with its circular shape and crescentic fold, the North Carolina specimen is a close match to those described

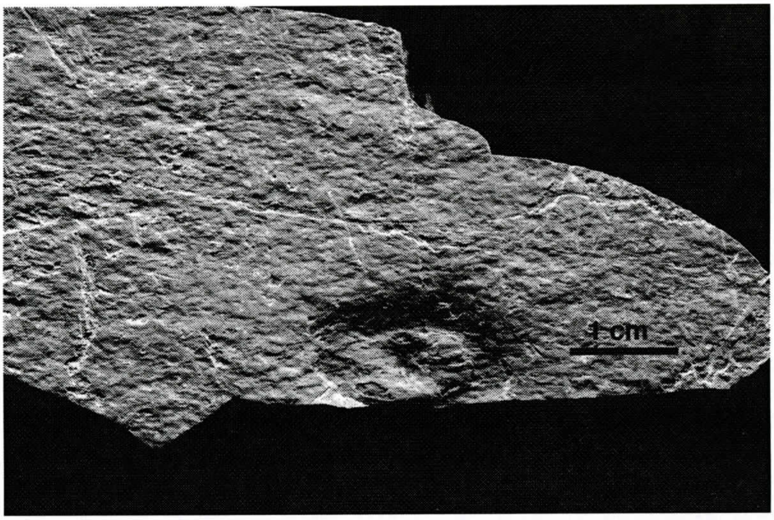



**Figure 4. *Aspidella* sp. cf. *A. terranovica* Billings 1872, NCSM 9713 recovered from locality 3 of Figure 2.**

from Sekwi Brook, Mackenzie Mountains by Hofmann (1981) as *Sekwia excentrica* and to a specimen of *Sekwia* described from the Clemente Formation of Sonora, Mexico (McMenamin 1996). Both the North Carolina and the Sonoran specimens occur on the base of sand beds, and both have identical diameters and comparable vertical relief. The new find thus adds considerable confidence to the validity of this Proterozoic genus.

McMenamin (2001) had transferred the Sonoran *Sekwia* to the species *Beltanelloides sorichevae* Sokolov, 1972. However, the clear appearance of the diagnostic crescentic fold in the North Carolina specimen suggests that *Sekwia* is a valid genus in its own right and not merely a preservational variant of *Beltanelloides*. Both the Sonoran and North Carolina specimens are hereby assigned to the species *Sekwia excentrica* Hofmann, 1981.

## *Aspidella* sp. cf. *A. terranovica* Billings, 1872 (Figure 4)

*Material—* Several specimens, the best of which, NCSM 9713, is preserved as a flat morph similar to Gehling et al, 2000 text figure 5.

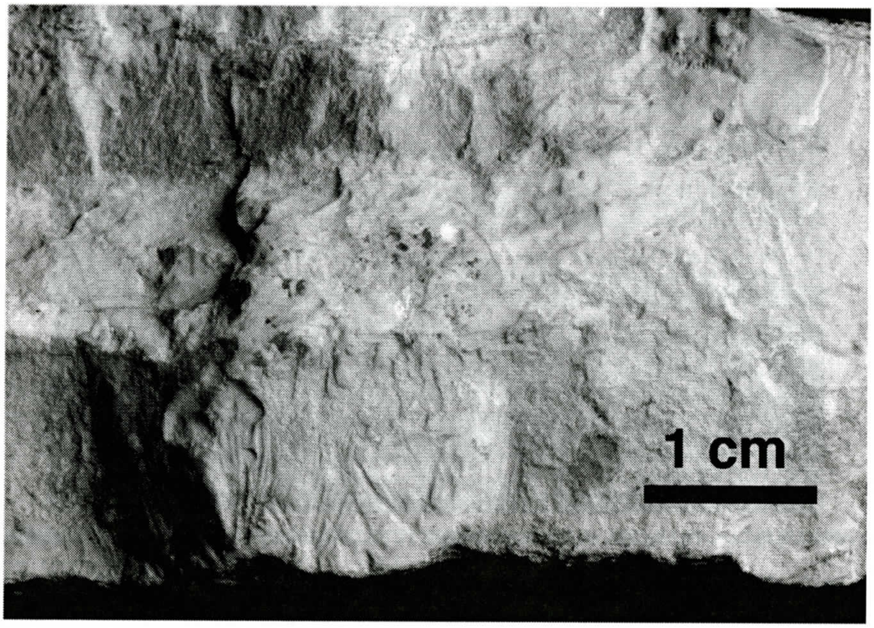
*Description—* The best specimen, NCSM 9713, is preserved as a flat to slightly concave elliptical disc, with a central concavity less than one quarter of the diameter of the whole disc. Between the rim and the central concavity, the disc is either smooth or ornamented with faint concentric rings near the outer rim.

*Locality—* Jacob's Creek Stone Quarry, near Denton, Davidson County, North Carolina. Locality 3 of Figure 2, unnamed Mudstone Member of the Cid Formation.

*Remarks—* This isolated specimen is preserved in a light gray, thinly bedded siltstone. The elliptical shape of the fossils is most likely the result of regional strain, as the trace of cleavage on bedding is parallel to the long axis of the *Aspidella* ellipses. *Aspidella* is thought to represent an "isolated holdfast of perhaps more than one genus of frond" (Narbonne et al., 2001), and similar considerations may apply to the holdfast fossil *Skewia*.

## Indeterminate Sac-Like Organism (Figure 5)

*Material—* one specimen has been recovered on a single float sample (NCSM 8714). The fossil on the rock occurs within several 10-14 mm thick tan to reddish tan mudstone layers. The upper surface of the fossil is truncated by erosion at the top of each layer. The specimen is a sac-shaped specimen 23 mm in width.



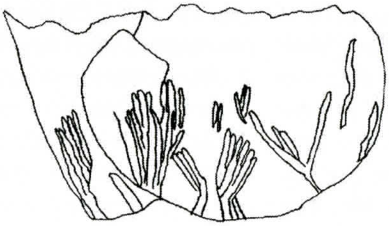



**Figure 5. Indeterminate sac-like organism, with line drawing to highlight the fossil, NCSM 9714 recovered from locality 2 of figure 2.**

*Description*— Laterally compressed sac-shaped fossil, consisting of a relatively smooth to slightly bulging wall grooved by intersecting sets of sub-parallel to fanned-out grooves. The grooves form an ornament on the sac surface that, along with the edges of the sac, runs at a high angle to bedding.

The grooves are 0.2-0.6 mm in width; where grooves are closely spaced and parallel. There may be a very narrow groove atop the ridge separating the two major grooves.

*Locality*— On private property southeast of Cottonville (near Mount Zion Church), Stanly County, North Carolina, Locality 1 of Figure 2, unnamed Mudstone Member of the Cid Formation.

*Remarks*— The specimen is 23 mm in width and 14 mm in height, but it appears truncated at the top of the bed that contains it so its original height is estimated to have been approximately 30-35 mm. It is possible that this specimen represents a ripped up piece of microbial mat. However as the specimen is contained in a finely bedded siltstone with no evidence that the bedding has been disturbed, it is more likely that this is a sac-like body fossil. Our earlier, tentative identification (Weaver et al. 2006b) of this specimen as ?*Inaria* Gehling, 1988 has been questioned by authorities on this genus (J. Gehling, personal communication, 2006), but nevertheless an attribution to this genus cannot be definitely ruled out at this point.

## CONCLUSIONS

There has been considerable discussion as to what factors (paleoenvironment, paleobiogeography, taphonomy, time) might influence assemblage types during the Late Neoproterozoic (Waggoner, 1999, 2003; Narbonne, 2005). Though the fossil record of Carolina Terrane Ediacara biota is sparse, it may give some insight into this on-going discussion. Due to the presence of *Pteridinium*, cf. ?*Swartpuntia*, true trace fossils and the age of the sediments in which these fossils were contained, this biota most closely matches what has been described by Waggoner (1999, 2003) as Nama Assemblage. As the sediments that contain Carolina Terrane Ediacara fossils are considered to have been deposited in much deeper water (Seilacher et al., 2005) than other Nama Assemblages, it is possible that time has a much stronger influence on Ediacaran assemblages than environmental factors.

## ACKNOWLEDGEMENTS

Authors are extremely grateful to North Carolina Fossil Club members Tony C. Furr and Ruffin Tucker for providing specimens for this study, and to Charles Brown, Charles Brown Photography for photographs of the specimens. We are also grateful to Janet Edgerton for providing bibliographic assistance and to Tamara Moore for assistance with Figure 2. Authors are also grateful to Dr. James Gehling and another anonymous reviewer for their insightful comments on an earlier draft of this manuscript and to Dr. Phil Novack-Gottshall and the editors of Southeastern Geology for their assistance with this version of this manuscript.

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V. 45, No.4, August 2008, p. 233-243

# A LATE PALEOCENE TEMPESTITE DEPOSIT EXPOSED ALONG A SECTION OF HURRICANE CREEK, DALE COUNTY, ALABAMA

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## ABSTRACT

**A layer of shell debris is exposed in the Tuscahoma Sand Formation (late Paleocene) along a portion of Hurricane Creek in Dale County, Alabama. The fossiliferous deposit contains an abundance of invertebrate gastropod and pelecypod shells derived from mixed environments ranging from estuarine to more nearshore open water settings. The morphology and condition of the shell-rich zone suggests that it formed through storm activity rather than through either condensed section accumulation or the bioturbation of sediment. The shells are normally graded and abraded or broken as a result of transport. Based on the geometry of the fossiliferous layer and the character and condition of the shells, it is proposed that the deposit was emplaced in a preexisting estuarine channel during the landward movement of a large tropical storm. As such, it represents a tempestite deposit. The shell unit exhibits a sharp contact with both under- and overlying lutite layers suggestive of rapid burial and little to no post-storm reworking. While the occurrence of a late Paleocene tropical storm is not unexpected for the northern Gulf Coastal Plain, the content and volume of the invertebrate fill material exposed along the creek bank at this locale indicates that the storm was both large and powerful. This tempestite provides a brief glimpse into shallow marine conditions during the late Paleocene and perhaps serves as an historic analog to possible conditions found in similar nearshore marine settings today.**

## INTRODUCTION

Fossiliferous layers of invertebrate shells generally form in sedimentary sequences by three different means: 1) through storm activity, 2) in sediment-starved areas via biological accumulation, and 3) due to the bioturbation of sediments and movement of the individual shells by trace makers. An understanding of the differences between these processes and the resulting shell bed morphology can aid in discerning the geologic history for many fossiliferous exposures.

A concentrated layer of invertebrate shells is exposed along a portion of Hurricane Creek in Dale County, Alabama. This fossiliferous layer is in the Tuscahoma Sand Formation, a late Paleocene unit in the Gulf Coastal Plain – Wilcox Group (Mancini and Tew, 1988, 1990). While this fossiliferous deposit has been briefly described in the geologic literature (e.g., MacNeil, 1946; Hastings & Toulmin, 1963; Toulmin, 1977), no analysis has been conducted at this locale to explain how this shell-rich unit may have formed or why it occurs between two dense lutite layers.

An investigation was undertaken to determine the manner by which this shell layer was formed and to study its stratigraphic relationship to adjacent beds. The presence of paleontologic materials in the Tuscahoma Sand, especially in southeastern Alabama, is uncommon and it could yield geological information relevant to understanding the former shallow marine setting for this locale during the late Paleocene.

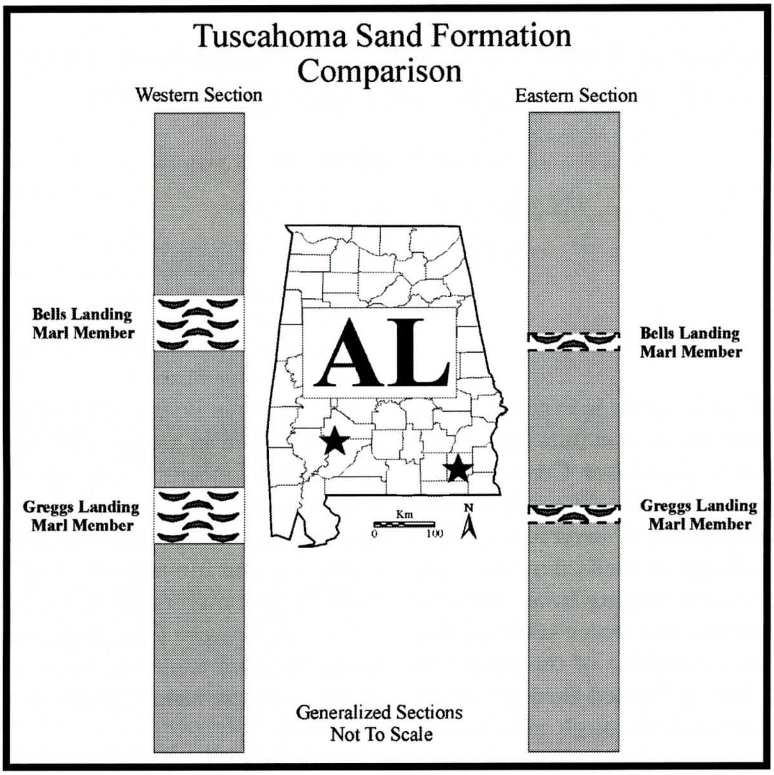


**Figure 1. Generalized composite Tuscahoma Sand Formation stratigraphic sections from Monroe County (west) and Dale County (east), Alabama. The Hurricane Creek locale in Dale County, (indicated by the black star) only exposes the Bells Landing Marl Member bounded by silty-clays. The Greggs Landing Marl Member has been reported along the Chattahoochee River. Not to scale and from various references (Adams and others, 1926; Toulmin, 1977; Raymond and others, 1988).**

## THE TUSCAHOMA SAND AND BELLS LANDING MARL MEMBER TYPE SECTIONS IN WESTERN ALABAMA

The Tuscahoma Sand type section is located in western Alabama along the Alabama River. It ranges in thickness from 76 to 107 meters, consists predominately of nonfossiliferous laminated silty clay and silt interbedded with fine-grained sand, and contains three or more fossiliferous glauconitic sand and marl zones in the lower half of the formation (Copeland, 1968). Two of the fossiliferous layers are given member status. The Bells Landing Marl Member occurs higher in the section and is approximately three meters thick. Located lower in the Tuscahoma Sand is the Greggs Landing Marl Member which is approximately two meters thick (Raymond and others, 1988) [Figure 1].

Additional exposures of the Tuscahoma Sand occur along the Tombigbee River (located west of the Alabama River) and these sections have also been used to describe the stratigraphy of this unit. For example, Toulmin (1977, p. 105) described the paleontology of the fossiliferous Bells Landing Marl Member exposed along the Tombigbee as a:

> ... bed containing abundant shells, the most abundant species being *Ostrea sinuosa*, *Venericardia aposmithii* Gardner and Bowles, and *Turritella postmortoni* Harris.

These invertebrate shells are an important component in defining the Tuscahoma Sand Formation, and specifically the Bells Landing Marl Member across the Alabama Gulf Coastal Plain.

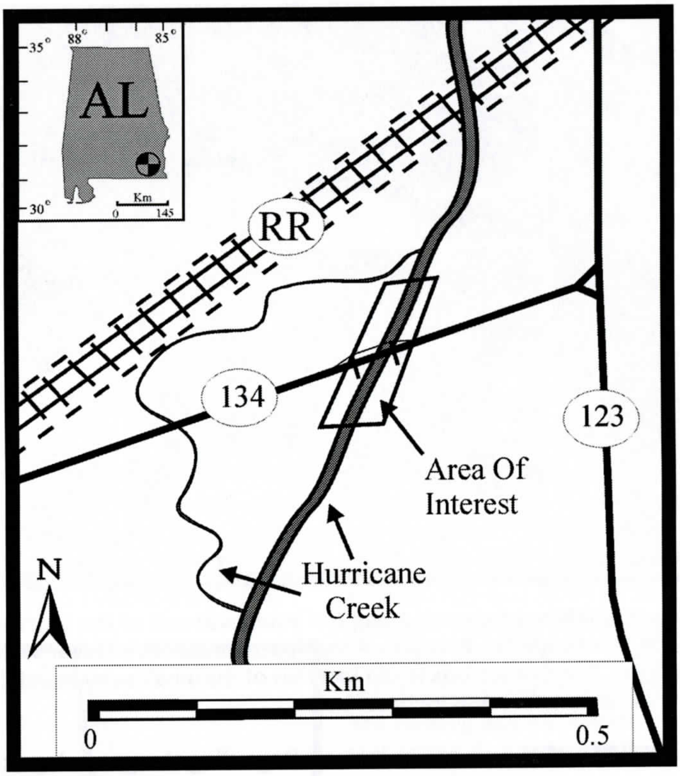


**Figure 2. The straightened portion of Hurricane Creek is the location of this investigation. The shell-rich layer outcrops on both sides of the creek in various places and is believed to represent a storm layer (i.e., tempestite) that has filled a former estuarine channel. The eastern side of the creek bank provides the best outcrops of the fossiliferous layer.**

## THE TUSCAHOMA SAND AND BELLS LANDING MARL MEMBER IN EASTERN ALABAMA

Geological exploration in eastern Alabama also identified the Tuscahoma Sand along the Chattahoochee River (Langdon, 1891; Smith and others, 1894; Toulmin & LaMoreaux, 1963; Toulmin and others, 1964). The Bells Landing Marl Member is identified along the course of the river by abundant *Ostrea sinuosa* and *Venericardia planicosta*, both very large and with other associated shells being decomposed beyond recognition (Smith and others, 1894). However, the Tuscahoma Sand is largely devoid of macro-invertebrate fossils in southeastern Alabama, with the exception of a few isolated locations. The most significant outcrop of the fossiliferous Bells Landing Marl Member occurs along a portion of Hurricane Creek in Dale County, Alabama.

## HURRICANE CREEK GEOLOGIC SETTING

### Location

Hurricane Creek flows predominately from the northeast to southwest and drains a small watershed located southeast of Ozark, Alabama. The area of interest is a straightened segment of the creek crossed by the Alabama Highway 134 bridge (Figure 2). The most extensive exposure of the shell bed is along the eastern side of the creek bank (Figure 3), a portion of which is covered by a geofabric and limestone riprap for erosional protection of the bridge.

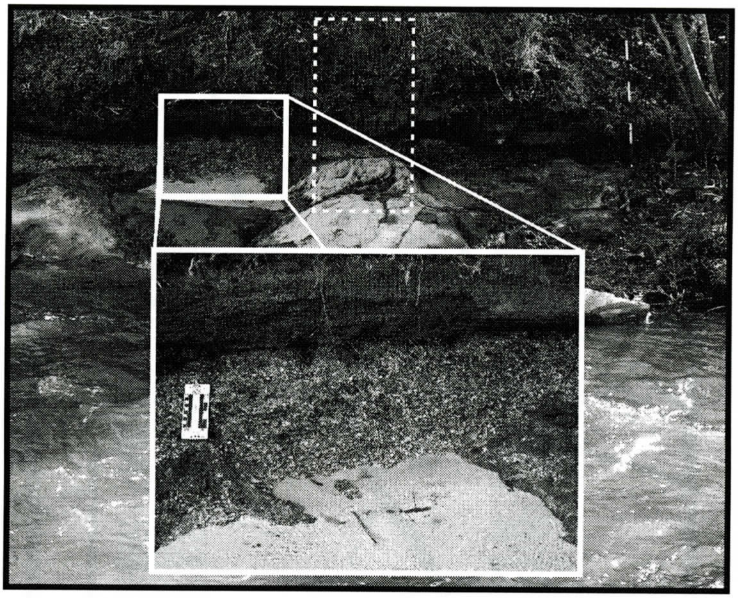

**Figure 3. A section of shell bed exposed along the southeast end of the outcrop. Scale on right side of outcrop is in 15-centimeter divisions. Panel insert presents a closer view of the normally graded shell-rich layer. The dashed box is the location of the small section exhibited in Figure 7. The insert scale is in centimeters and inches.**

## Stratigraphy

Working in southeast Alabama, Smith (2001) defined the Tuscahoma Sand as:

> ... about 80 to 125 feet [24 to 38 m] thick and generally consists of a thin basal glauconitic sand overlain by dark-gray to black, thinly laminated, micaceous and carbonaceous, nonfossiliferous clay and silty clay. (p. 79) [Brackets mine]
>
> The Tuscahoma represents a regressive or southward shift of the shoreline, with fine lignitic and carbonaceous clay accumulation in restricted or closed bay or lagoon environments under lowered salinities and reducing conditions. The uniform lithology of these beds indicates relatively constant environmental conditions with little or no influx of coarse clastic sediment into the area. The more or less constant and uniform environment further suggests equilibrium between the rate of sea-level fall, or decreased rate of subsidence, and influx of sediment. (p. 80)

Regarding the general strike and dip of the Tuscahoma Sand, Smith (2001, p. 80) wrote:

> In central and eastern Dale County, ...the Tuscahoma Sand strikes northeast to east-northeast and dips toward the southeast at about 25 feet per mile [4.7 m per kilometer]. [Brackets mine]

## Paleontology

MacNeil (1946) was the first to report a significant outcrop of Tuscahoma Sand macro-invertebrate fossils along Hurricane Creek. In describing the fossiliferous layer he stated:

> This deposit near Newton in Dale County, appears to be of shallow-water or beach origin, and consists mostly of water worn shells. The deposit is roughly equivalent to the Bells Landing Marl Member of the Tuscahoma of western Alabama and may be at the same stratigraphic position, but no exposures are known between this locality and central Wilcox County. (p. 19)

Several years later, this same locale was visited by Hastings & Toulmin (1963). They defined the shell-rich layer as:

> The exposure, of Tuscahoma Sand, consists of gray firm silty fine-grained sand and gray thin bedded silty clay or shale. A discontinuous irregular fossiliferous layer consisting largely of more or less worn and abraided (sic) shells and phosphorite pebbles indicate current-style deposition in a near-shore area. Some shells are little worn and must have been transported a short distance. The fossils are similar to those in the Bells Landing Marl Member near the middle of the Tuscahoma Sand in exposures on the Alabama River. The most common and conspicuous fossils are large *Ostrea compressirostra* [i.e., *Ostrea sinuosa* Rogers and Rogers], *Venericardia apos-mithii*, *Turritella postmortoni*, and the colonial coral *Haimesiastraea conferta*. (Hastings & Toulmin, 1963, p. 17) [Brackets mine]

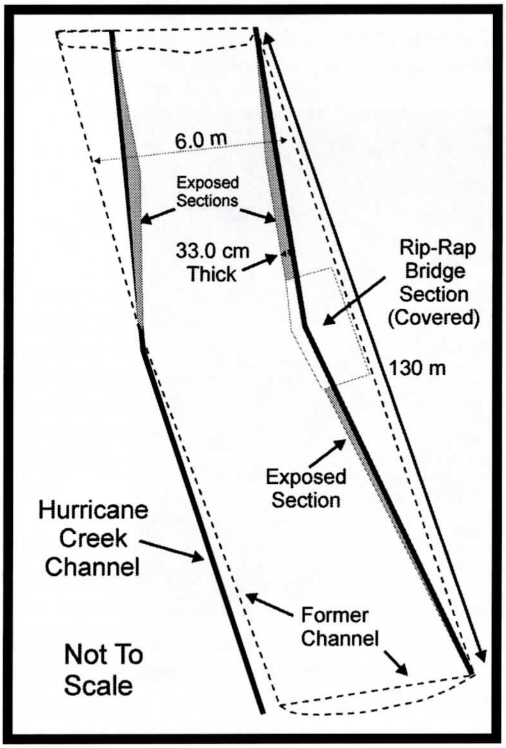


**Figure 4. The general dimensions of the former shell-filled channel are projected along Hurricane Creek. The configuration of the former channel is estimated from the shell-bed outcrops that extend along portions of both creek banks. Drawing not to scale.**

## FIELD OBSERVATIONS AND ANALYSIS

The area of interest occurs along a straightened portion of Hurricane Creek where the stream cuts through the former shell-filled channel (Figure 4). The shell bed is exposed along the eastern creek bank for a distance of approximately 130 m. This is not the limit of the fossiliferous unit as it likely extends further northward within the subsurface for an unknown distance. The southernmost exposure of the fossiliferous layer eventually terminates downstream as a result of the Hurricane Creek channel exceeding the dimensions of the smaller shell-filled channel. The straightened channel cuts somewhat diagonally through the shell bed making any estimation of the size and extent of the fossiliferous layer somewhat problematic. However, based on shell bed exposures along both creek banks, the former channel is estimated as having been 6.0 m wide by 33.0 cm deep. Shell thickness within the former channel varies as a result of the undulating bottom surface and also due to pinch and swell in the two underlying lutite layers.

The shell bed channel floor is covered in places by large (up to 15 cm in diameter) unbroken *Ostrea sinuosa* shells. The lower section of the shell bed has greater concentrations of smooth-shaped oblong silty-clay clasts (some up to 5.0 cm in length) and large rounded pieces of broken *Ostrea sinuosa* (some measured up to 7.6 cm in length) [Figure 5]. *Venericardia* shells also occur in the lower section of the biogenic layer. The remaining shell fill is dominated by *Turritella* shells, some up to 3.8 cm in length along with some smaller rounded clay clasts. Shell abrasion of the innumerous *Turritella* is consistent throughout the entire outcrop. The fossiliferous deposit exhibits normal graded bedding in a silty-clay matrix and can best be defined as a biogenic packstone (Dunham, 1962; Nichols, 1999). The top of the fossilifer-

**Figure 5. A slab of shell bed that displays the underside of the fossiliferous layer. Note the large pieces of oyster (*Ostrea*) and clam (*Venericardia)* shells and rounded clay clasts. Large and dense materials compose the lower portion of the shell layer. Scale in centimeters and inches.**

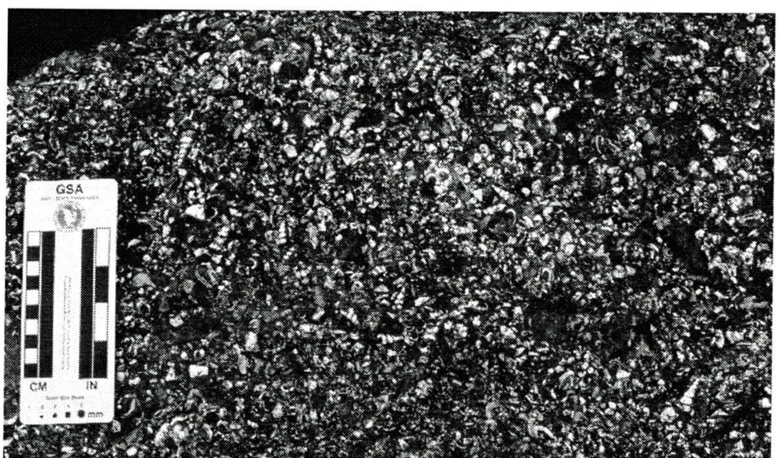

**Figure 6. The top of the fossiliferous layer. Note the abundance of *Turritella* and small clay clasts. The excellent preservation of the top of the shell unit reflects rapid burial and little to no post-storm alteration of the storm deposit. Scale in centimeters and inches.**

ous layer exhibits no shell maceration which suggests rapid burial (Alexandersson, 1979) [Figure 6]. Intermixed within the shell layer is sparse previously unreported abraded vertebrate material including fish vertebrae, shark teeth, and pieces of stingray dental plate. The poor condition and limited number of these fossils does not allow for specific identification (i.e., genus/species) of these vertebrates.

Color variations occur in several of the clay

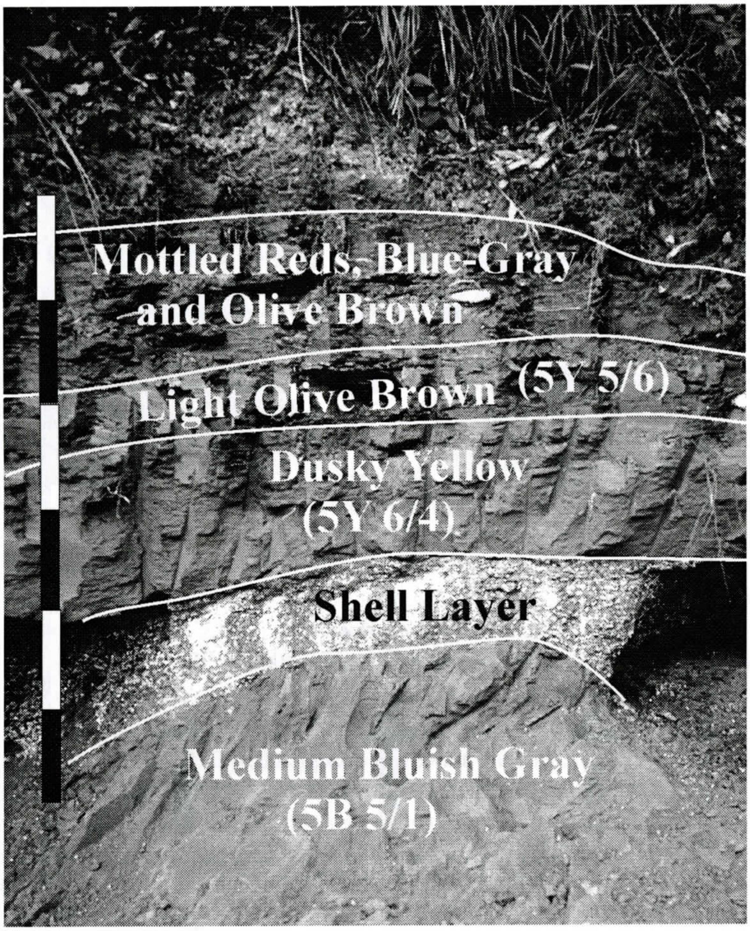


**Figure 7. A section of the Tuscahoma Sand Formation exposed along Hurricane Creek. The shell layer (i.e., tempestite) is bounded by clays of different colors. The shell layer is pinched at this spot (compare with Figure 3). There are no obvious physical characteristics between the clay layers (other than color) that would allow one to determine which layer or layers might be associated with the shell layer as a storm deposit. Likely some of the overlying clay was storm generated, but its differentiation is not possible by examination with a hand lens. Scale in 15-centimeter divisions.**

layers (Figure 7). However, no visible differences in clay content other than color is discernable by inspection using a hand lens. Any attempt to designate one or more of the overlying lutite layers to the tempestite would be subjective as the color differences appear to be diagenetic. None of these clay layers were examined in an attempt to identify any microfossils.

## METHODS OF SHELL CONCENTRATION

Some shell concentrations are proposed to have formed in sediment-starved areas due to a decrease in clastic sedimentation and the development of extensive faunal communities on or within the existing substrate (Fürsich, 1978; Kidwell 1982, 1985, 1986, 1991; Kidwell and others, 1986). This setting is typically associated with a transgressive to highstand stratigraphic sequence along with the development of a condensed section (Kidwell, 1982; Banerjee &

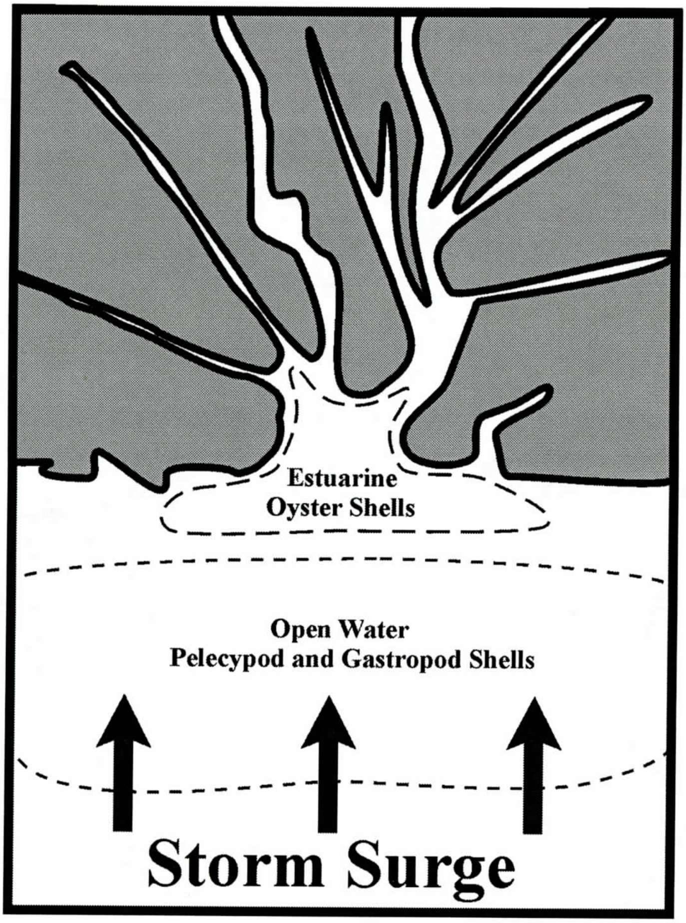


**Figure 8. Conceptualized drawing of the envisioned former paleosetting that was impacted by a large and powerful tropical storm. The fossiliferous deposit found along Hurricane Creek suggests that open water gastropod and pelecypod shells were transported landward with the storm surge. These open water shells were abraded and in many cases broken during transport. The shells were deposited on top of estuarine oyster shells and mixed with rounded lutite clasts in a former estuarine paleochannel.**

Kidwell, 1991).

Another means by which shells can become concentrated is through biological sediment-mixing activities. Trace makers can modify substrate sediments through bioturbation thereby concentrating the shells within a clearly defined zone within the subsurface (Schäfer, 1952; van Straaten, 1952, 1956; Rhoads & Stanley, 1965; Frey, 1973; Meldahl, 1987). Multiple burial events could then correspond to numerous layers of concentrated shells within a vertical section.

The occurrence of shell accumulations has also been attributed to the effects of storm events. Shells are winnowed from the substrate, transported some distance, and then deposited in varying concentrations in subaqueous environments or washed onshore (Hayes, 1967; Andrews, 1970; Kumar & Sanders, 1976; Kreisa & Bambach, 1982; Aigner, 1985; Morton, 1988; Davis and others, 1989; Isphording & Isphording, 1991; Froede, 2005).

## DISCUSSION

The fossiliferous layer exhibited along a portion of Hurricane Creek has physical characteristics typical of a sedimentary deposit generated by a large storm event and not by processes more commonly attributed to either sediment-starvation or concentration through bioturbation. Additionally, a tempestite deposit documenting the passage of a tropical storm across the submerged Gulf Coastal Plain during the late Paleocene is not unexpected.

The lutite strata that over- and underlie the fossiliferous layer appear to have been deposited in a low-energy estuarine subtidal marine setting (Figure 8).

There is no evidence of bioturbation beneath the fossiliferous layer along the former channel. The presence of oblong and rounded lutite rip-up clasts of various sizes mixed throughout the fossiliferous bed suggests that some level of channel scouring probably occurred during the course of the storm. The combination of rip-up clasts and shell debris indicates that there was a significant level of storm energy associated with the formation of this shell-rich deposit.

The normal grading of the shell bed suggests a decrease in shoreward-directed transport energy as the channel filled with storm-derived debris. It also implies that the channel served as a conduit in the movement of water during the course of the storm. This interpretation is also supported by the variation in levels of shell abrasion that occur in the various invertebrate species. In many places unbroken *Ostrea* shells line portions of the bottom of the paleochannel. Their excellent condition indicates a nearby source area and limited transport distance. Above these shells are the broken *Ostrea* fragments and large rounded oblong clay clasts, indicating greater transport (and abrasion) distance. Above this are greater concentrations of well-worn *Turritella* mixed with far fewer and smaller clay clasts. The *Turritella* shells equate to a nearshore subtidal inner-shelf setting which suggests they were derived from a more open water environment and transported a greater distance shoreward.

The contact between the top of the fossiliferous layer and the overlying silty-clay layer is abrupt. There is no indication of any post-storm erosion, dissolution, or bioturbation of the shell material. Perhaps a post-storm change in water depth or an abrupt increase in the sedimentation rate restricted the alteration, bioturbation, or reworking of the shell-rich storm layer. While one or more of the overlying clay layers might be attributable to the same storm event that created the tempestite bed, the stratigraphic location at which to mark this contact would be arbitrary. The silty-clays are of a similar composition and the color changes are likely diagenetic.

## CONCLUSION

While many different interpretations of this shell-rich deposit are possible, the morphology, scale, bedding, and stratigraphic position of the fossiliferous layer between dense silty-clay layers suggests that it formed from a large and powerful tropical storm. Gastropod and various pelecypod shells winnowed from an open water setting were transported landward and deposited into a preexisting estuarine channel. Sufficient burial occurred following the storm to prevent the reworking or loss of this storm deposit from the geological record. This locally-restricted high-energy event deposit provides a snap shot of the marine conditions within southeastern Alabama during this portion of the late Paleocene and might prove useful as a historic analog to conditions likely found in similar marine settings today.

## ACKNOWLEDGMENTS

Gratitude is expressed to Mr. A. Jerry Akridge and Dr. William J. Neal for their helpful and constructive comments. Mrs. Thajura Harmon-Unongo provided valuable reference assistance. Dr. Charles C. Smith kindly supplied materials related to his geologic work in southeastern Alabama. This work neither represents the views or opinions of the U.S. Environmental Protection Agency, nor was this investigation conducted in any official capacity. Any mistakes that may remain are the responsibility of the author.

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SOUTHEASTERN GEOLOGY
V. 45, No.4, August 2008, p. 245-254

# EVALUATION OF CLAY MINERALOGY AS A SEDIMENT PROVENANCE INDICATOR IN VIRGINIA

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## ABSTRACT

**Samples of clay-rich alluvium from the banks of low-order streams in small, single- or simple-lithology watersheds in central and northwestern Virginia were analyzed to determine the extent to which sediment clay mineralogy reflects its parent material. Bedrock types include shale, limestone, biotite gneiss, schist, slate, diorite, and hornblende metagabbro. Powder x-ray diffraction of clay separates and bulk sediment was performed to identify minerals present. Petrographic examination of sediment and parent rock samples was also used to identify the mineralogy of the silt- and sand-sized sediment fractions. All alluvium samples contain kaolinite + hydroxy-interlayered vermiculite ± illite. Samples can be classified as (a) illite-dominated and shale-related or (b) kaolinite-dominated (with or without illite) and related to any of the other studied rock types. These simple clay mineralogical suites reflect surficial erosion of residual soils and are not sufficiently distinctive to be a useful provenance indicators for alluvial materials in the old and deeply weathered landscape of Virginia. Mineralogy of the sand-sized fraction of alluvium, while invariably quartz-rich, is a closer reflection of bedrock type.**

## INTRODUCTION

The issue of sediment provenance is one of interest to many earth scientists (Pettijohn, 1975). For example, environmental geologists who study questions of geomorphology and land use effects may be interested in the origins of sediment transported by a stream so that areas of intense erosion can be pinpointed or sediment budgets can be constructed (Youngberg and others, 1975; Yee and Parsons, 1980; Meade, 1982; Hsieh, 1984; Motha and others, 2003). The key to provenance determination is finding characteristic(s) of each source area that are distinctive enough leave a signature on sediment transported out of the area. The means by which provenance is determined varies depending on the nature and origin of the sediment but often involves its mineralogy, chemistry, or related characteristics (Klages and Hsieh, 1975; Wall and Wilding, 1976; Oldfield and others, 1979; Walling and others, 1993; Hillier, 2001; Motha and others, 2003).

While sand provenance has been extensively investigated by sedimentary petrologists (Pettijohn, 1975; Dickinson and Valloni, 1980; Zuffa, 1985; Dorsey, 1988), the geographic origins of finer grained sediment is often of more use to environmental geologists. Clays and silts can be transported easily to streams by sheetwash and other soil erosion processes. These small grain sizes are also more likely than sand to travel as suspended load, rather than bedload, contributing to stream turbidity and being more easily sampled.

Clay mineralogy of alluvial sediment has been investigated for a variety of reasons by different workers, and some regions have been found to have distinctive clay-mineral signatures (e.g., Potter and others, 1975; Ambers, 2001). In some of these areas, clay mineralogy has been successfully used as a provenance tool (Klages and Hsieh, 1975; Hsieh, 1984; Moslehuddin and others, 1998; Valero-Garces and others, 1999). One commonality among these areas is that they have a sufficiently wide variety of bedrock types for weathering processes to result in different clay minerals.

Previous work on residual soils in Virginia, a state with a diversity of rock types and ages, found variation in clay mineral suites formed on different bedrock types (e.g., Plaster and Sherwood, 1971). The purpose of this study is to investigate the possibility of using clay mineralogy for provenance determination of alluvium in Virginia. Small, low-order watersheds, most of which have only a single bedrock lithology, were targeted for sampling; and a wide range of bedrock types were included in the study. Alluvium samples were collected from stream banks at 11 different sites and analyzed for mineralogy. The hypothesis that clay mineralogy can be used as a provenance tool was then tested by determining the degree to which different bedrock types produced alluvium with a distinctive clay-mineral signature.

## METHODS

Various geologic maps of the central Virginia Piedmont and northwestern Virginia Valley and Ridge provinces were consulted to locate small, low-order, single-lithology watersheds in which to sample (Rader, 1967; Ern, 1968; Spencer, 1968; Brown, 1969; Rader and Evans, 1993; Virginia Division of Mineral Resources, 1993). Only one of the sampled sites (site 11) had a watershed with mix of rock types, and these were all related sedimentary rocks dominated by shale.

Residual soils in the sampled watersheds are predominantly deeply weathered Ultisols with some Inceptisols in higher relief areas (Natural Resources Conservation Service, 2007). Sampled alluvium was presumed to be derived from this weathered residuum originating from bedrock within modern drainage basin boundaries. Most likely the age of these samples is recent, as thick deposits of historical alluvium cover valley bottoms throughout this region as a result of soil erosion caused by agricultural practices of European-American settlers (Trimble, 1974; Ambers and others, 2006).

A global positioning system (GPS) receiver was used to record the location of each of the 11 sampling sites, all of which were stream bank exposures. Relatively cohesive, clayey deposits were targeted. If multiple layers of alluvium were observed in the bank cut, or if relatively homogeneous sediment had color differences due to oxidation/reduction, each major layer was sampled, for a total of 21 sediment samples. Samples of the local bedrock, in the form of relatively unweathered stream cobbles, were obtained at each site where available. Sediment samples were placed in resealable plastic bags and taken back to the laboratory for analysis.

Particle-size analysis was accomplished with sieves and hydrometer using ASTM method D422 (American Society for Testing and Materials, 2003). After the last hydrometer measurement for each sample, supernatant liquid containing the clay-sized fraction was poured off and the clays were allowed to settle out. Clay samples were then prepared for x-ray diffraction using the Millipore® filter transfer method described by Moore and Reynolds (1997) to produce oriented mounts. An Epsom salt solution was used to saturate the clays on the filter cake with $Mg^{2+}$. A second oriented mount saturated with $K^+$ was prepared using a KCl solution. Randomly oriented bulk mounts of moist sediment (intact, not size fractionated) were also made.

X-ray diffraction was performed using a Rigaku Miniflex+ Portable/Desktop X-ray Diffraction System with copper K$\alpha$ radiation and a nickel monochromator. Using a step size of 0.02 and a count time of 1 second, clay samples were run from 2° to 30° 2$\theta$ and bulk sediment samples were run from 2° to 50° 2$\theta$. Each $Mg^{2+}$-saturated clay sample was run three times: air-dried, glycerol solvated, and heated to 550°C for at least an hour. Each $K^+$-saturated sample was run once (air-dried), as were the bulk sediment samples. After x-ray diffraction was complete, Jade software and Chen (1977) were used to identify mineral peaks on the diffractograms. NEWMOD© was used to generate clay mineral patterns (Reynolds, 1985; Walker, 1993), then a graphical NEWMOD© file-mixing program called MODBLEND was used to model the diffraction patterns to obtain semi-quantitative proportions of the different clays present in each sample.

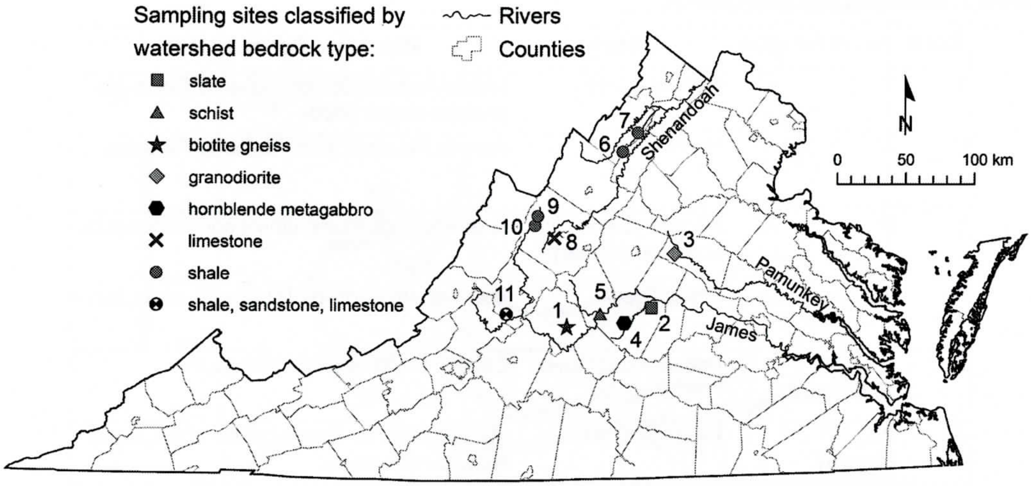


**Figure 1. Map of Virginia showing numbered sampling site locations and bedrock types in stream watersheds.**

Intact pieces of 13 of the sediment samples were air-dried then oven-dried at 60°C and vacuum-impregnated with epoxy. These were sawn and sent off for thin sectioning, along with chips of the bedrock samples. A Nikon E400 petrographic microscope was later used to examine the resulting thin sections and make qualitative observations of mineralogy and textural relationships.

## RESULTS

The 11 stream bank sampling sites are shown in Figure 1 and described in Table 1. Watersheds with metamorphic and igneous bedrock types are located in the Piedmont, and watersheds with sedimentary bedrock are located in the Valley and Ridge province. All but three of the samples are from alluvial deposits taken from 0.3 to 2 m depth along stream cut banks. One sample along a first-order stream bank (site 5) was from a residual B-C horizon of soil and another (site 10) was colluvial, as no alluvial deposits could be located at these sites. One of the three samples taken from site 6 was clay from the stream bed, which appeared to be highly weathered shale bedrock.

Particle-size analysis of the samples reveals that despite their cohesive feel, only one sample contains >30% clay (Figure 2). Clay contents typically range from 9-30%; sand and silt contents range from 44-77% and 37-68%, respectively. The one sample (from site 8) that contains 46% clay is unusual in its color and consistency, being waxy, sticky and so rich in very fine organic matter that it is almost black. In the field, this material is overlain by reddish-brown silty clay more typical of the other deposits sampled. It is likely that the black silty clay was deposited in a transient lacustrine environment such as a beaver pond or oxbow lake.

X-ray diffraction of the clay-sized fraction of the samples reveals a relatively simple mineralogy. All the samples contain a little quartz and goethite. Clay minerals include only kaolinite, illite, and a dioctahedral form of vermiculite known as hydroxy-interlayered vermiculite (HIV) (Barnhisel and Bertsch, 1989). Figure 3 shows an example set of diffractograms for a sample from site 4.

The relative proportions of these three clay minerals vary depending on the bedrock type of the watershed from which the samples were obtained (Figure 4). Samples from streams draining biotite gneiss and slate, along with one of the two diorite samples, do not contain illite and are dominated by kaolinite with minor HIV (up to 12%). Samples from areas underlain by schist, hornblende metagabbro, and limestone, and the other diorite sample contain mainly ka-

**Table 1. Sampling site information.**

| Site # | No. of Samples | Stream | Watershed Bedrock Units |
|---|---|---|---|
| 1 | 2 | Rutledge Creek | Middle Proterozoic meta-igneous biotite-plagioclase augen gneiss |
| 2 | 2 | Hunts Creek | Arvonia Fm slate, some Buffards Fm schist |
| 3 | 2 | Wheeler Creek unnamed tributary | Green Springs Pluton diorite and hornblendite |
| 4 | 2 | Forsip Creek | Buckingham Complex hornblende metagabbro |
| 5 | 1 (B-C horizon) | Branch Pond Creek | Candler Fm mica schist and phyllite |
| 6 | 2 + bed clay | Passage Creek | Marcellus Shale, Needmore Fm shale, some Ridgeley Sandstone |
| 7 | 2 | Passage Creek unnamed tributary | Mahantango Fm, Marcellus Shale, Needmore Fm shale, some Ridgeley Sandstone |
| 8 | 2 | Eidson Creek | Conococheague Fm and Beekmantown Fm limestones and dolostones |
| 9 | 1 | Braley Branch | Brallier Fm shale and sandstone, some Millboro Shale and Needmore Fm shale |
| 10 | 1 (colluvium) | West Dry Branch unnamed tributary | Brallier Fm shale and sandstone, some Chemung Fm sandstone and shale |
| 11 | 3 | Cedar Creek | Millboro Shale and Needmore Fm shale, some Edinburg Fm limestone and Conasauga Shale and related Lower Devonian and Upper Silurian sedimentary rocks |

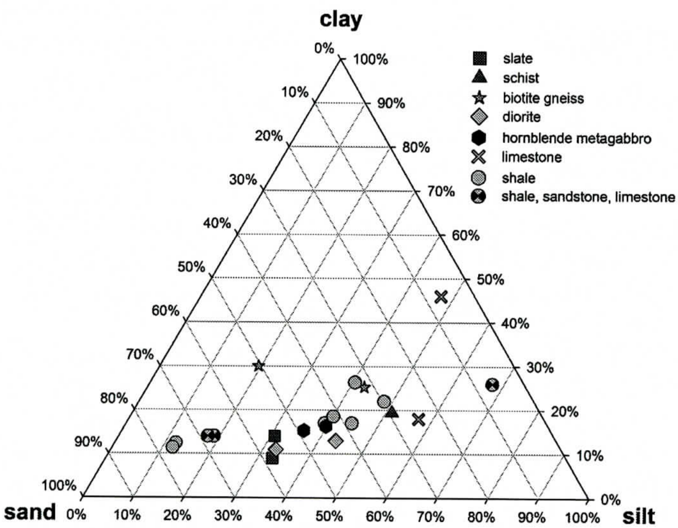


**Figure 2. Textural triangle showing the percentages of sand, silt, and <2-μm clay in each sample, classified according to watershed bedrock type.**

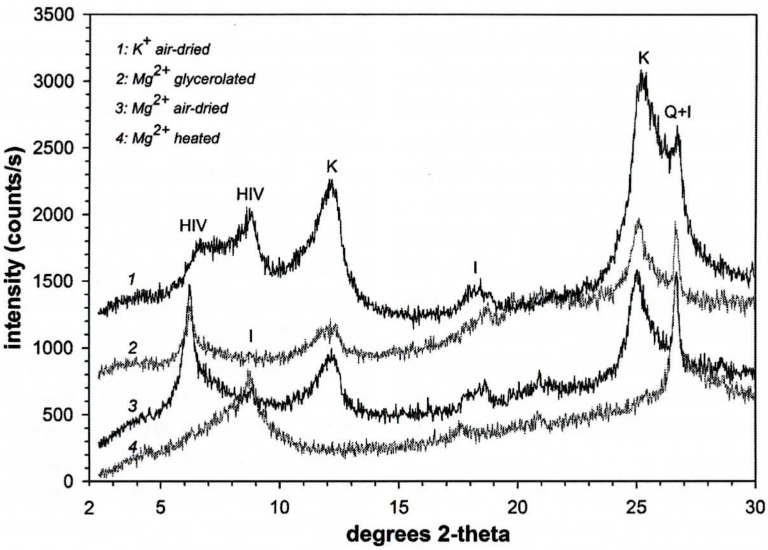


**Figure 3. X-ray diffractograms of different treatments on a sample from site 4. The intensity of the $K^+$ air-dried, $Mg^{2+}$ glycerolated and heated patterns have been changed by 750, 500, and -225 counts, respectively, to separate the patterns for greater visibility. Peak labels: K = kaolinite, I = illite, HIV = hydroxy-interlayered vermiculite, Q = quartz.**

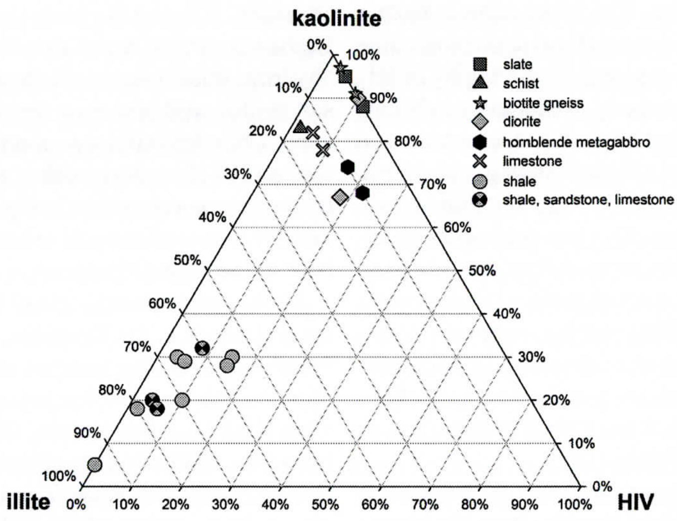


**Figure 4. Ternary diagram showing the percentages of the three main clay minerals found in alluvium samples classified according to watershed bedrock type. All the samples with >60% kaolinite are located east of the Blue Ridge except those from a limestone watershed. Shale-dominated watersheds produce alluvium with >50% illite.**

olinite with up to 22% HIV and 15% illite. In these two groups, the only samples from west of the Blue Ridge are those from the limestone watershed (site 8). The rest are derived from igneous and metamorphic rocks of the Piedmont (sites 1-5). All the other samples are from shale or shale-dominated watersheds in the Valley and Ridge and contain predominantly illite with significant kaolinite (18-32%) and minor HIV (2-15%). The one sample that contains 95% illite and no HIV is highly weathered shale from the stream bed at site 6.

Petrography shows that sand mineralogy is much more varied than that of the clay fraction of sediment samples and more closely resembles bedrock mineralogy. Quartz is ubiquitous and feldspars are common in alluvium samples, but hornblende is only found in sediment from the hornblende metagabbro and diorite (sites 3, 4). Biotite is found in samples from the biotite gneiss and diorite watersheds (sites 1, 3); and muscovite is found in samples from the slate, schist, and biotite gneiss watersheds (sites 1, 2, 5). Lithic fragments of bedrock are present in alluvium from the limestone, slate, schist, and shale-dominated watersheds (sites 2, 5, 8, 11).

## DISCUSSION

Multiple samples taken at the same stream location show consistent clay mineralogy at all sites except in the diorite watershed (site 3) and one shale watershed (site 6). In the diorite case, the upper sample from a reddish, oxidized zone contains 15% illite. The lower sample from a dark gray, reduced zone closer to the water line contains no illite apparent on the $Mg^{2+}$- or $K^{+}$-saturated diffractograms. Reducing conditions due to water saturation in the stream bank are unlikely to degrade illite (Fanning and others, 1989). It is possible, however, that mineralogical variation in this alluvium represents an inverted soil profile formed from progressive erosion and deposition of surface (illite-absent) then subsurface (illite-bearing) residual soil horizons. Illite content of residual soils can increase with depth as weathering in surface horizons converts it to other minerals such as HIV (Allen and Hajek, 1989).

At site 6, the stream bed clay is almost pure illite, whereas alluvial samples from the stream banks contain approximately 15% HIV and 30% kaolinite. These differences can be attributed to formation of the bed clay from *in situ* weathering of shale. The alluvial bank material is derived instead from soils formed on this bedrock type. More intense leaching in the soil environment provides opportunities for formation of HIV and kaolinite (Allen and Hajek, 1989).

The alluvium samples in this study can be separated into two main groups: illite-dominated and kaolinite-dominated (Figure 4). The kaolinite-dominated samples can also be split into those which do and do not contain illite, although this trait is not entirely site-dependent, as the diorite samples illustrate.

The dominance of kaolinite in Piedmont metamorphic- and igneous-derived alluvium versus illite dominance in Valley and Ridge sedimentary-derived alluvium is consistent for all samples except those from site 8. This site in the Shenandoah Valley has limestone bedrock in its watershed but kaolinite dominance. The texture of the two samples from the site is different, one being quite clay- and organic-rich (Figure 2); but their clay mineralogy is very similar in containing kaolinite >> illite > HIV (Figure 4).

Illite and kaolinite are both common clay minerals in soils formed on limestone (Dixon, 1989; Fanning and others, 1989). In a study of soil developed on the Lenoir limestone (now known as the Lincolnshire limestone) near Staunton, VA, not far from site 8, Carroll and Hathaway (1953) found that illite (they term it "hydrous mica") was most abundant deep in the soil profile and absent in the A and A-B horizons, while kaolinite was abundant throughout the profile. Assuming such a pattern holds for the limestone soils of the sampled watershed, it implies that surficial soil erosion provided the bulk of the alluvial sediment.

Studies of sediments carried by two large rivers in the region, the Rappahannock and James, found the same clay mineral suite of kaolinite, illite, and HIV as did this investigation of alluvium (Nelson, 1958; Nelson, 1962; Feuillet and Fleischer, 1980). Studies of the region's residual soils, however, have found more complex clay-mineral suites often including minerals such as smectite, chlorite, and gibbsite (Cady, 1951; Rolfe, 1953; Plaster and Sherwood, 1971; Harris and others, 1985; Graham and others, 1989; Jolicoeur and others, 2000). Where found, these minerals tend to occur deep in the soil profile and alter to HIV and/or kaolinite in the more heavily leached, acidic surface horizons (Borchardt, 1989).

The absence of smectite in all alluvium samples studied here is particularly intriguing, as smectite is common in the alluvium and suspended sediments of other regions such as the Mississippi River valley (Johns and Grim, 1958; Johnson and Kelley, 1984). Bedrock types in central Virginia known to produce smectitic soils, such as the hornblende metagabbro, were specifically targeted in this study in an attempt to find mineralogical variation. Assuming that smectite is present in soils of at least some of the sampled watersheds (as Plaster and Sherwood, 1971, suggest), its absence implies, once again, that sediments were primarily derived from erosion of surface soil horizons that lack smectite.

Alternatively, conversion of smectite to HIV and kaolinite might be taking place fairly rapidly within alluvial deposits due to acidic, leaching conditions in these loamy, well-drained soils. The typically very fine particle size of smectite (Borchardt, 1989) could also enable it to remain suspended in flood waters and largely avoid deposition with coarser clays, silts, and sands; but this seems improbable in overbank events in which floodwaters soak into and dry up on the floodplain. Should smectite or other deep-profile, easily altered minerals such as chlorite or gibbsite be found in other studies of alluvial deposits in this region, their presence would most likely indicate deep soil erosion in the watershed, recent deposition, and/or poor drainage conditions in the alluvium.

In terms of the usefulness of clay mineralogy as a provenance tool for alluvium, the results of this study show that there is not much hope for this method. While there is variation in the relative amounts of kaolinite and illite, both kaolinite and HIV are ubiquitous (if only present in very small amounts in some alluvium samples) and no more distinctive minerals are present in the clay fraction. Bedrock lithologies as diverse as metagabbro, mica schist, diorite, and limestone produce virtually indistinguishable alluvial clay-mineral suites.

The best case scenario for provenance determination would be one in which a stream bearing kaolinite-dominated sediment flowed into an illite-dominated stream, resulting in an intermediate amount of kaolinite and illite downstream. The proportions of the two minerals might then be used to determine the relative contributions of the two streams to the suspended sediment load below the confluence. The spatial distribution of the two mineralogical groups makes this scenario likely only in the Valley and Ridge where limestone and shale watersheds may be adjacent to one another.

Another possibility would be a large river like the James that drains both the Valley and Ridge and the Piedmont provinces. A study of its suspended clay mineralogy could potentially reveal the relative contribution of fine sediment from the two provinces. The spatial resolution of sediment sources could not be much finer than province-level, however, which severely limits the usefulness of the technique.

The problem for clay provenance studies in Virginia is that the majority of clays that end up in alluvial deposits appear to derive from erosion of the surface horizons of soils. Under the prevailing climatic conditions and vegetation of this region, these horizons are acidic and heavily leached, resulting in alteration of most clays to HIV and/or kaolinite with some illite remaining, depending on how illite-rich the source rock is. Deep soil erosion that would tap into horizons containing more distinctive, bedrock-related minerals such as smectite and chlorite seems relatively uncommon, at least in the small watersheds studied here. This is encouraging as far as watershed system health goes but not very useful for provenance analysis.

A better option for provenance determination of alluvium would be the sand fraction of sediment. Quartz is ubiquitous; but many more distinctive, unaltered minerals and lithic fragments may be present in this size range. In the samples observed here, many quartz grains were rather coarse, so sieving out the coarsest fraction of sand may concentrate the more useful non-quartz mineral grains. Quantification of mineral abundance would then require point counting sediment thin sections manually or using automated tools.

The drawback for environmental geologists studying sediment actively transported by a stream is that sand is less likely than clays to be carried as wash load in small streams. It is more likely to be part of the bed material load, sometimes traveling suspended near the bottom of the water column and other times traveling with the bed load (Gordon and others, 2004). This means that acquiring sand to study may necessitate bed load as well as integrated suspended sediment sampling.

## CONCLUSIONS

The clay mineralogy of alluvium derived from residual soils formed on bedrock types including shale, limestone, slate, schist, biotite gneiss, hornblende metagabbro, and diorite is very simple. Illite-dominated samples associated with shale can be distinguished from kaolinite-dominated samples (with and without illite) from all the other rock types, but no other consistent distinctions can be made. HIV is the only other clay mineral found, and it is ubiquitous in amounts ranging up to one-quarter of the clays present. The reason for this relatively uniform mineralogy is that most alluvial clays appear to be derived from erosion of highly weathered surface soil horizons, rather than deeper horizons where more bedrock type-dependent clay minerals may be found.

Clay mineralogy is therefore not highly useful for provenance determination of alluvium or suspended sediments in the parts of central and northwestern Virginia investigated here. Given the similar climate and vegetation conditions throughout much of the Southeast, clay provenance is not likely to be of great utility anywhere in the region except in locations where one or more clay minerals other than kaolinite, HIV, and illite are sufficiently abundant in surficial soil layers. Sand, being less weathered and more closely tied to bedrock mineralogy, may be a better size fraction to study for environmental geologists in need of sediment source information.

## ACKNOWLEDGMENTS

This material is based upon work supported by the National Science Foundation under Grant No. EAR-0420317 and by grant J-755 from the Thomas F. and Kate Miller Jeffress Memorial Trust of Virginia. I would like to thank undergraduates Melanie B. Stine and Sara R. Rothamel for summer laboratory and field assistance, Dr. Clifford P. Ambers for sound advice throughout this project, and Dr. Cullen Sherwood for a helpful review of the manuscript.

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